Representation of Stormflow and a More Responsive Water Table in a TOPMODEL-Based Hydrology Model

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Abstract

This study presents two new modeling strategies. First, a methodology for representing the physical process of stormflow within a TOPMODEL framework is developed. In using this approach, discharge at quickflow time scales is simulated and a fuller depiction of hydrologic activity is brought about. Discharge of water from the vadose zone is permitted in a physically realistic manner without *a priori* assumption of the level within the soil column at which stormflow saturation can take place. Determination of the stormflow contribution to discharge is made using the equation for groundwater flow. No new parameters are needed. Instead, regions of near saturation that develop during storm events, producing vertical recharge, are allowed to contribute to soil column discharge. These stormflow contributions to river runoff, as for groundwater flow contributions, are a function of catchment topography and local hydraulic conductivity at the depth of these regions of near saturation. The second

approach improves groundwater flow response through a reduction of porosity and field capacity with depth in the soil column. Large storm events are better captured and a more dynamic water table develops with application of this modified soil column profile (MSCP). The MSCP predominantly reflects soil depth differences in upland and lowland regions of a watershed. Combined, these two approaches—stormflow and the MSCP—provide a more accurate representation of the time scales at which soil column discharge responds and a more complete depiction of hydrologic activity. Storm events large and small are better simulated, and some of the biases previously evident in TOPMODEL simulations are reduced.

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Introduction

Successful modeling of the hydrologic cycle requires representation and quantification of the various pathways by which water migrates through a catchment. Over the last twenty years, the capabilities and fidelity of hydrology models have increased considerably. Many factors impacting the spatial and temporal variability of the hydrologic cycle, and the scales at which these processes operate, have been elucidated and applied to model simulations. These developments have permitted modeling of the spatial distribution of soil moisture levels across the land surface, water movement within the soil column, and the relative contributions of evapotranspiration and river runoff to the efflux of water from the catchment system.

Many hydrology models have been developed using TOPMODEL formulations (Ambroise et al., 1996a; Beven and Kirkby, 1979; Koster et al., 2000; Sivapalan et al., 1987; Stieglitz et al., 1997). A framework of analytic equations, TOPMODEL is based on the idea that topography is the primary determinant of the distribution of soil moisture at and within the land surface (Beven, 1986a; Beven, 1986b; Beven and Kirkby, 1979; Beven et al., 1994).

TOPMODEL formulations define areas of hydrological similarity—that is, points within a watershed that respond to meteorological forcing in similar fashion, saturating to the same extent, producing the same levels of discharge, etc. These points of hydrological similarity are identified by an index that is derived from analysis of catchment topography. This topographic index is often of the form $ln(a/tan\beta)$, where $tan\beta$ is the local slope angle at a patch on the land surface, and a is the amount of upslope area draining through that patch. Lowland areas tend toward higher topographic index values, due to a combination of either low slope angle or large upslope area. Upland areas tend conversely toward lower topographic index values. Points within a catchment with the same topographic index value are assumed to respond identically to atmospheric forcing. Thus within a TOPMODEL framework, the topographic index provides the fundamental unit of hydrological response.

This fundamental topographic unit is derived from three basic assumptions (see Ambroise et al., 1996a; Beven, 1997 for details):(1) the water table is approximately parallel to the topographic surface so that the local hydraulic gradient is close to $tan\beta$; (2) the saturated hydraulic conductivity falls off exponentially with depth; and (3) the water table is recharged at a spatially uniform, steady rate that is slow enough, relative to the response time scale of the watershed, to allow the assumption of a water table distribution that is always at equilibrium. These assumptions permit reconstruction of the spatial variability of catchment response to meteorological forcing solely from modeling of the response of the mean state. This quasi-stochastic approach is at once computationally efficient while still permitting dynamic representations of physical processes within the system.

These conceptual underpinnings provide a foundation for physically-based modeling of catchment hydrology. Land surface models based on these formulations have been applied to catchments both large (Ducharne et al., 2000) and small (Beven and Kirkby, 1979). However, while the TOPMODEL framework has provided hydrologists with a powerful and efficient tool for modeling hydrologic conditions, the full complement of dynamic processes is not represented in most TOPMODEL applications. As a consequence, model simulations often perform poorly during drier conditions. The response of watersheds to wetting by spring snowmelt and to storms after an extended dry period have proven particularly difficult to represent. Correction of these simulation inaccuracies requires a more detailed depiction of the hydrologic cycle and catchment physical structure.

Discharge response

Four physical processes contribute to river runoff in a watershed: 1) precipitation onto stream channels; 2) overland flow; 3) shallow subsurface stormflow; and 4) groundwater flow (Hornberger et al., 1998). The first two of these processes respond very rapidly, producing spikes in hydrographs during and immediately after storm events.

The third mechanism, shallow subsurface stormflow, responds at the quickflow time scale. (N.b. In this paper the terms stormflow and groundwater flow refer to physical processes; the terms quickflow and baseflow define short and long time scales of response.) Tracer studies have shown that shallow, subsurface regions of the soil column can support significant levels of flow during storm events (Dewalle and Pionke, 1994; Hendershot et al., 1992; Ogunkoya and Jenkins, 1993). Such regions can exist as perched water tables, disconnected from the true water table supporting groundwater flow (Gile, 1958; Hammermeister et al., 1982; Noguchi et al., 1999; Wilson et al., 1990). These perched water tables, by virtue of their development in the vadose zone nearer the land surface, can flow more quickly, discharging their waters more rapidly to the catchment river network. However, the timing of the development of these perched water tables and their size and placement within the soil column have proven difficult to model.

Stormflow has been represented in Variable Infiltration Capacity (VIC) models (Liang et al., 1994; Lohmann et al., 1998a; Lohmann et al., 1998b). VIC models provide a viable modeling alternative to the more physically-based TOPMODEL approach. These parameterized reservoir models mimic the time scales of catchment hydrologic response to storm events and can be structured to emulate quickflow and baseflow time scales. Such time scales are calibrated to inferred rates of quickflow and baseflow derived from runoff analyses.

Recently, Scanlon et al. (2000) included stormflow in a TOPMODEL-based hydrology model through introduction of a parameterized quickflow reservoir. This model represents a hybridization of the parameterized reservoir and TOPMODEL approaches. The authors used a saturation deficit model as their basic soil column structure, then added a second saturation deficit reservoir from which stormflow would be determined. The groundwater flow and stormflow components were partitioned so as not to overlap and thus supersaturate the soil column. Analyses of hydrograph response to different storm events and antecedent conditions, and piezometer sampling of the study catchment were used to determine the saturation deficit recession coefficients. Two time scales of response, an order of magnitude apart—471 and 36 hours for groundwater flow and stormflow respectively—were delineated; these rates were applied to the reservoir discharge formulations. The authors used a linear rate of decrease in the transmissivity of the soil column for the stormflow reservoir, while maintaining an exponential decay of transmissivity for the groundwater reservoir. Flow from the subsurface stormflow reservoir to the groundwater reservoir was facilitated by a linear recharge function. For the study watershed, the model depicted both the rapid and slow discharge of water from the soil column following storm events.

This hybrid modeling approach (Scanlon et al., 2000) produced three additional parameters: one for stormflow reservoir maximum capacity; a second for the stormflow discharge rate; and a third for the linear recharge function. Successful simulation of catchment hydrology required calibration of these parameters to rates inferred from analysis at a highly instrumented experimental watershed. An attractive alternative to this hybrid model approach would be one allowing a more general inclusion of stormflow, while requiring fewer parameterizations. Indeed, if stormflow could be incorporated in a manner consistent with the TOPMODEL formulations that govern the flow of water within the soil column, no additional parameters would be necessary.

The fourth process contributing to river runoff in a watershed, groundwater flow, provides most of the baseflow during extended periods between storm events. This discharge mechanism is represented in land surface hydrology models constructed in the TOPMODEL framework (Ambroise et al., 1996a; Ambroise et al., 1996b; Stieglitz et al., 1997; Wood et al., 1990). Specifically, TOPMODEL formulations permit dynamically consistent calculations of both the partial contributing area—from which precipitation onto stream channels and overland flow can be determined—and the groundwater flow that supports this area. However, model recession hydrographs often run high during wet periods, then low during drier months, as will be shown in this paper. Consequently, calibrations applied during wet conditions corrupt simulation accuracy during dry conditions, and vice versa.

These model discharge biases reflect inappropriate levels of groundwater flow generation. Within the TOPMODEL framework, groundwater discharge is determined by the height of the water table (or analogously by the saturation deficit). Low groundwater flow simulation during dry conditions stems from greater depth of the water table and concomitant lower hydraulic conductivities controlling both discharge and recharge rates. During such dry conditions the model water table height often is not sufficiently responsive to recharge, and thus groundwater flow discharge is muted. A more responsive water table needs to be represented if groundwater discharge rates are to be more accurately modeled.

Such a dynamic water table could be effected by increasing infiltration-recharge rates; however, these rates are set by Darcy's Law and the TOPMODEL assumption that saturated hydraulic conductivity decays exponentially with depth—the same attributes controlling groundwater discharge. Thus, recharge cannot be altered without changing groundwater discharge, or decoupling the mechanisms controlling these two processes—an unrealistic anisotropy. Instead, a physically representative modification is needed that engenders greater response of the water table to recharge, thereby intensifying groundwater flow response.

Here we present two new strategies intended to address these model shortcomings: 1) a physically-based approach to modeling stormflow and its application in a manner consistent with TOPMODEL assumptions; 2) a modified soil column framework, in which porosity and field capacity are realistically allowed to change with depth, that provides a more responsive water table and better groundwater flow simulation. Together, these strategies produce a more complete and accurate depiction of hydrologic activity at the catchment scale.

The Hydrology Model

The hydrology model employed for this study has been previously described (Stieglitz et al., 1997). Two methods are used for modeling the flow of water within a catchment. The first is a soil column model that simulates the vertical movement of water and heat within the soil and between the soil surface plus vegetation to the atmosphere. The ground scheme consists of ten soil layers. Layer thicknesses are structured in a geometric series determined from the depth of the first ground layer—typically 4 centimeters for this study. Diffusion and a modified tipping bucket model govern heat and water flow, respectively. The prognostic variables, heat and water content, are updated at each time step. In turn, the fraction of ice and temperature of a layer may be determined from these variables. Transpiration and other surface energy balance calculations use a standard vegetation model (Pitman et al., 1991) that includes bare soil evaporation and canopy interception loss.

The second method partitions the catchment surface into two distinct hydrologic zones: saturated lowlands; and unsaturated uplands. Using the statistics of the topography, the horizontal movement of groundwater is tracked from the uplands to the lowlands (a TOPMODEL approach). Combining these two approaches (Figure 1) produces a three-dimensional picture of soil moisture distribution within a catchment. The partitioning of runoff and surface water and energy fluxes is effected without the need to explicitly model the landscape. Specifically, an analytic relation, derived from TOPMODEL assumptions, exists between the mean water table depth $(\bar{z})$, determined from the soil column model, and local water table depth at any location x (z_x) (Sivapalan et al., 1987; Wood et al., 1990)

$$z_x = \bar{z} - 1/f\left[ln(a/tan\beta)_x - \lambda\right] \qquad (1)$$

where $ln(a/tan\beta)_x$ is the local topographic index at location x; λ is the mean watershed value of $ln(a/tan\beta)$, and f is the rate of decline of the saturated hydraulic conductivity with depth in the soil column. By setting z_x equal to zero, i.e., locating the local water table depth at the surface, saturated regions of the land surface can be identified. This partial contributing area includes all locations for which

$$ln(a/tan\beta)_x \geq \lambda + fz \qquad (2)$$

From this partitioning, the contributions to river runoff of both precipitation directly onto stream channels and overland flow (saturation-excess runoff) can be quantified. Following Sivapalan et al., (1987) groundwater flow (Q_b) is:

$$Q_b = \frac{AK_s(z=0)}{f} e^{-\lambda} e^{-fz} \qquad (3)$$

where A is the area of the watershed, and K_s is the saturated hydraulic conductivity at the surface. This flow through the soil matrix supports river discharge between storm events.

This combined approach to modeling the land surface has been validated at several watersheds, ranging in scale from 2.2 km^2 (Stieglitz et al., 1999) to 570,000 km^2 (Ducharne et al., 2000).

The Approach

A Conceptualization of Stormflow

Within our hydrology model, Darcian flow accounts for recharge of the water table and occurs within layers of the soil column for which volumetric soil moisture is greater than 70% of field capacity (see Hillel, 1977). Our conceptualization of stormflow makes use of this condition. We argue that not only will Darcian flow produce vertical recharge of the water table, but it will also bring about stormflow discharge. Such stormflow is a natural consequence of the topographic variability of the watershed and the TOPMODEL assumption that water tables form parallel to the land surface. While our hydrology model conceptualizes a vertical soil column and horizontal land surface, in fact most of the land surface is sloped (Figure 2). Thus, gravity will not force Darcian flow within the soil column solely in the direction of the water table—a component of the flow will be directed laterally. Just as the true water table migrates through the soil column, regions of near saturation in the vadose zone will also produce discharge. It is thus possible to develop a representation of shallow subsurface stormflow making use of the existing formulations for groundwater flow (i.e. Equation 3). As with groundwater flow, gravity guides the motions of these stormflow waters.

Model stormflow is initiated within a layer of the vadose zone when the volumetric soil moisture level exceeds 70% of the field capacity (Figure 3). Calculation of the stormflow component begins with redistribution of the excess water into a fully saturated region from the base of the layer upward:

$$sw_i = zb_i - \left(\frac{\theta_i - 0.7\theta_{33i}}{\phi - 0.7\theta_{33i}} \right) \Delta z_i \qquad \theta_i > 0.7\theta_{33i.} \qquad (4)$$

where sw_i is the stormflow water table in layer i, zb_i is the bottom boundary of model layer i, θ_i is the volumetric soil moisture in layer i, θ_{33i} is the field capacity in layer i, and Φ_i is the porosity in layer i. As in Stieglitz et al. (1997), groundwater flow is calculated given a water table of that height (Q_{Swi}). Groundwater flow is then calculated a second time for a water table of height to the top of the layer below (Q_{Sbi}). This second value is subtracted from the first, leaving only the flow within the layer in question.

$$Q_{Si} = Q_{Swi} - Q_{Sbi} \qquad \textbf{(5)}$$

Total stormflow for a given time step is merely the sum of the stormflows for each layer in the vadose zone.

$$Q_S = \Sigma Q_{Si} \qquad \textbf{(6)}$$

This formulation requires no new parameterizations. It incorporates the local hydraulic conductivities within each layer. Saturated zones that develop nearer to the surface flow more quickly than zones formed at depth. This circumstance necessarily generates different rates of stormflow and groundwater flow. These differences reflect the time scales of quickflow and baseflow.

A Modified Soil Column Profile (MSCP)

Catchment upland areas tend to possess shallower soils than lowland areas (Cox and McFarlane, 1995; Webb and Burgham, 1997; Yanagisawa and Fujita, 1999). Consequently, at the depth of upland bedrock, lowland areas may still hold sedimented sands, clays, and organics, supporting a greater porosity and field capacity. This has some important implications. When the catchment is very wet, a given amount of water table recharge is spread across the entire catchment, since the water table exists everywhere. When the catchment is dry, on the other hand, the water table is limited to the lowlands, and a given amount of recharge is allocated to only a fraction of the catchment. As a result, in the dry catchment, the level of the water table increases more quickly for a given amount of recharge -- the dry catchment has a more responsive water table.

A one-dimensional soil column framework, of course, can never allow the explicit treatment of this three-dimensional behavior. Nevertheless, a suitable parameterization could allow the one-dimensional model to capture this behavior in a gross sense. That is, a parameterization could be devised that makes the water table respond more dynamically to recharge as the level of the water table decreases.

We introduce such a parameterization here. Like many hydrology models, our starting model used a single parameter value for porosity and a second for field capacity, regardless of the depth of the soil column (Stieglitz et al., 1997). With the new parameterization, the imposed porosity and field capacity in the soil profile decrease with depth into the soil, so that a unit of recharge increases the water table level more quickly when the water table is low (that is, when the catchment is dry). Such a modified soil column profile (MSCP) is consistent with TOPMODEL steady-state assumptions. This MSCP could, in a sense, be interpreted as an "averaging" of soil depth properties over the upland and lowland areas of the catchment -- as the depth into the soil increases, the fraction of the catchment with bedrock at that depth increases, so that the average porosity decreases.

Porosity, in addition, has been shown to decrease with depth at a point, as demonstrated in field site analyses using a variety of methods, including mercury intrusion (Ajmone-Marsan et al., 1994), bulk density (Asare et al., 1999; Bonell et al., 1981; Cox and McFarlane, 1995), and fractal approaches (Bartoli et al., 1993; Oleschko et al., 2000). These profiles reflect near-surface bioturbation and soil compaction. Although the impact of compaction and bioturbation on water table dynamics is probably significantly less important than that of spatial variability in bedrock depth, the observed one-dimensional porosity profiles are at least consistent with the proposed parameterization.

The MSCP should produce a more responsive water table and generate greater groundwater flow. To test this hypothesis, we applied simple and equal geometric reductions to both the porosity (cm^3/cm^3) and the field capacity (cm^3/cm^3) – ranging from 1% to 15% per layer -- to the soil column of our hydrology model. By the above arguments, steeper catchments with shallower upslope soils should require a greater rate of decrease of porosity and field capacity with depth.

Model Simulations

Experimental watersheds at Sleepers River and at Black Rock Forest were used in this study. These sites are topographically representative of the rolling hillslope and steeper ravine catchments that dominate the hydrology of the northeastern United States. Meteorological conditions at both watersheds typify the daily, seasonal and interannual variability of this mid-latitude region. Periods of drought, particularly during summer, are not uncommon for either watershed.

At each study site, four model simulations were performed: 1) a baseline run without any of the new modifications; 2) simulation with stormflow; 3) simulations with the MSCPs; 4) simulations with both stormflow and the MSCPs.

Results

Application to the Sleepers River Watershed

The Sleepers River watershed (111 km^2) located in the glaciated highlands of Vermont is hydrologically representative of most upland regions in the northeast United States. As such, this site was chosen in 1957 as an experimental watershed by the Agricultural Research Service (ARS) to provide a better understanding of natural watershed behaviour and aid in the development of testing physically based hydrologic models (Anderson, 1977). Nested entirely within the Sleepers River Watershed is the W-3 sub watershed (8.4 km^2). The topography is characterized by rolling hills, and the soils are predominantly silty loams. Vegetation cover is approximately equally distributed amongst grasses, coniferous forest and deciduous forest. Hourly measurements of air temperature, dewpoint temperature, incoming shortwave and thermal radiation and wind speed are recorded within the watershed. Average annual air temperature is 4.1°C with a standard deviation of 11.4°C. Mean hourly precipitation is determined from 7 gauges placed within the W3-sub-catchment; annual precipitation totals approximately 109cm.

Another data set contains the snow water equivalent (SWE), snow depth, snow temperature and soil temperature. Total snow depth averages about 254cm and snow cover usually persists from December to March with snowmelt in late March and April. Hourly runoff data are available from a gauge at which the Pope Brook leaves the W3-sub-catchment of the Sleepers River (Stieglitz et al. 1997). Five years of meteorological and hydrologic data collected between 1969-74 were used to drive the model.

Hydrographs of the Sleepers River model runs are shown for years 1971 and 1973 (Figures 4 and 5). The baseline model runs depict the gross characteristics of runoff generation; however, many simulation inaccuracies are apparent (Figures 4a and 5a). Specifically, we identify three model shortcomings (see Figure 4a):

1) Model response to storm events during dry conditions—i.e. October through April—is often reduced and spiky, lacking a short-term discharge recession curve in the days immediately following a storm.
2) Groundwater flow response is high during transitions from wet to dry conditions—such as the long May through July 1971 recession at Sleepers River.
3) Model response to initial wetting from spring snowmelt is muted and delayed.

A fourth shortcoming, low bias between storm events during dry periods, is also apparent. However, most likely this bias reflects the absence of any depiction of deep aquifer water discharge. Such deep flows operate at very long time scales, which are not represented in TOPMODEL dynamics. It is not our intention in this study to address this shortcoming.

With stormflow activated there is improvement in the match of modeled and measured hydrographs (Figures 4b and 5b). This model simulation generates river discharge at shorter time scales, better representing storm event response, especially in dry conditions. Smaller storm events are now resolved, and recession curves after summer storm events are evident. There is also considerable improvement in the model hydrograph response to spring snowmelt.

Model simulations with the different MSCPs produce modest improvement in the match of modeled and measured hydrographs. A reduction in porosity and field capacity of 5% was found to best improve model discharge (Figures 4c and 5c); however, this simulation still fails to capture many smaller storm events and for those it does, produces little short-term recession in the days immediately following the storm. At the baseflow time scale, however, groundwater flow is more responsive, producing greater total discharge than the baseline run. Consequently, the May 1971 response is exaggerated, but the long July recession following this event is better matched (Figure 4c).

Sensitivity analyses were performed to determine which MSCP worked best in conjunction with stormflow. With both stormflow and a 5% reduction of porosity and field capacity, modeled and measured hydrographs match best (Figures 4d and 5d). Once again, the stormflow component generates discharge at quickflow time scales. The model is responsive in both wet and dry conditions, simulating discharge recession for storms large and small, and best captures catchment response to spring snowmelt. Dry season storm events are well depicted; however, groundwater discharge bias between these storms persists.

Application to the Black Rock Watershed

The Black Rock forest is a 1500 hectare preserve located in the Hudson Highlands region of New York. Elevations in the forest range from 110 to 450 m above mean sea level, with seasonal temperatures ranging from –2.7 °C to 23.4 °C. The medium texture soils are typically very thin, with parent material located from 0.25 to greater than 1m below the surface in the depressional areas. Soils in the lowland areas are more organic than upslope, but bulk densities are not significantly different. Exposed bedrock is common throughout the preserve and consequently the area was not extensively farmed during the period of European settlement. Lumber extraction ceased in 1927 and the forest has been managed as a preserve without significant disturbance since that time. The system is typical of the *Quercus* dominated, secondary growth forests that have characterized the NE United States over the past century. The catchment (1.35 km^2) is drained by a single stream, Cascade Brook. Average hourly discharge from Cascade Brook is monitored continuously using a V-notch weir installed in 1998. Hourly measurements of precipitation, air temperature, dewpoint temperature, incoming shortwave radiation and wind speed are also taken. Hourly thermal radiation for the site is calculated following the methodology of Anderson and Baker (1967). Three years of meteorological and hydrologic data, collected between 1998-2000, were used to drive the model. (Note that the Black Rock weir malfunctioned during November 1999.)

Hydrographs of the Black Rock model runs are shown for years 1999 and 2000 (Figures 6 and 7). The baseline simulations at Black Rock (Figures 6a and 7a) are considerably less accurate than their Sleepers River counterparts. This circumstance is due in part to differences in catchment structure. The Black Rock watershed is steeper, rockier and possesses shallower soils than Sleepers River. Two of the three shortcomings apparent in the Sleepers River baseline run are evident from the Black Rock baseline simulation:

1) Dry condition storm events are not depicted.
2) Groundwater discharge is high during transitions from wet to dry conditions.

Warm winter conditions precluded significant snowpack development at Black Rock from 1998-2000. Consequently, the third shortcoming, poor response to initial wetting from spring snowmelt, is absent. Of particular note is the long summer drought of 1999 that was broken in September by Hurricane Floyd (Figure 6a). The baseline model run fails to simulate this flood event, and the smaller storms subsequent.

The addition of stormflow generates a better discharge simulation (Figures 6b and 7b). Smaller storm events—particularly during drier periods—are better captured, and there is a visible short-term recession curve following the initial flood event. However, not all of the characteristics of the measured hydrographs are well matched. In particular, large storm events are still not well simulated, and local river runoff maxima, such as the overtopping of the weir during Hurricane Floyd in 1999 are still often underrepresented.

Model runs with the MSCPs produce considerable improvement of discharge simulation. A reduction in porosity and field capacity of 12% was found to best improve model discharge (Figures 6c and 7c). The shallow, rocky soils of hillslope and upland regions of Black Rock justify this larger rate of decrease in porosity and field capacity. The soil profile modification intensifies groundwater flow response to events poorly captured in the baseline simulation, producing greater total discharge. The initial flood due to Hurricane Floyd is now well depicted, as is the large storm in March 1999 (Figure 6c). Many smaller storm events, however, remain unresolved, and the MSCP model run simulates little of the short-term recession present in the measured hydrographs.

Analyses were again performed to determine which MSCP worked best in conjunction with stormflow. With both stormflow and a 12% reduction of porosity and field capacity, modeled and measured hydrographs match best at Black Rock (Figures 6d and 7d). Smaller storm events are resolved, and once again the stormflow component generates discharge at quickflow time scales. The model is responsive in both wet and dry conditions and best captures the intensity of catchment response to large storm events. Hurricane Floyd is well captured, as is the large storm in March 1999, and the recessions following these flood events are represented (Figure 6d). Some problems remain: groundwater flow response remains low between storm events during dry periods; and runoff generation is too heightened in late 2000.

Analysis of Water Table Depth

Further illustration of the impacts of stormflow and the MSCP can be seen from examination of modeled, mean water table depth, ($\bar{z}$). Figure 8 presents comparisons of the three new model simulations with the baseline run for the Sleepers River catchment.

With stormflow activated (Figure 8a) less water recharges the water table than for the baseline run. As a consequence, the water table is lower, and groundwater flow is necessarily diminished; however, rather than a reduction in hydrograph response, the missing water mass is instead shunted to stormflow discharge, raising river runoff levels following storm events (Figures 4b and 5b). Not only is the timing of discharge thus redistributed, but overall the catchment is more responsive with stormflow activated, generating greater total discharge for the five-year duration of the simulation (data not shown).

Model simulation with the MSCP—5% reduction in porosity and field capacity—produces a more dynamic water table (Figure 8b). The modified soil column holds less water at depth and therefore is more responsive to an equal volume of recharge. This greater responsiveness necessarily intensifies groundwater flow generation and shortens the time scales at which this discharge occurs. Additionally, the water table drops more abruptly in response to an equal volume of discharge. Thus, the water table rises higher in the soil column with wet conditions but lower during dry periods. These changes in water table response and groundwater flow production are not offsetting; rather, the raising of the water table during wet conditions has a disproportionately larger impact on discharge quantities than the lowering during dry periods. Overall, more total discharge is generated. This more dynamic water table explains the May 1971 response at Sleeper

River, in which peak runoff is overly simulated, but July recession is well matched (Figure 4c).

The effects of stormflow and the MSCP on water table depth offset one another during wet conditions (Figure 8c). This balance reduces the exaggerated discharge simulation found in the model run with the modified soil column profile alone; however, the improvement of the July 1971 recession at Sleepers River remains (Figure 4d). During dry conditions, the impacts of the two modifications are additive, as both bring about a lowering of the water table. As a consequence, underrepresentation of discharge persists between dry period storm events. In spite of this bias, water is still available for stormflow generation, and the recharge that does reach the water table can raise it more quickly, producing faster wetting of the catchment during spring snowmelt.

Impact of Soil Layer Structure

Additional analyses were performed to determine the sensitivity of the combined stormflow/MSCP simulations to the layer resolution of the model soil column. Layer thicknesses were altered by changing the depth of the first ground layer, from which all other layer depths are calculated (see Methodology). Runs with both stormflow and the MSCP were made at both Sleepers River and Black Rock with first ground layer depths of 2, 3, 4, 5, 7 and 12 centimeters. Discharge simulations were found to be insensitive to these alterations and therefore appear robust.

Spectral Analysis

Whereas the model hydrograph improvements effected by stormflow and the MSCP are readily apparent at Black Rock, the impacts at Sleepers River are more difficult to discern. To elucidate better the differences at Sleepers River, the faster time scales of discharge response were assessed by spectral analysis. Fourier analyses were performed on all five years of the measured runoff record and model simulations (January 1, 1970 through September 30, 1974). Figure 9 shows the full spectral profile of the measured runoff. Most of the power is in the low frequencies, matching the cycle of annual snowmelt at Sleepers River.

Figure 10 presents a partial spectral profile of the higher harmonics (200 and higher). The measured runoff record (Figure 10a) has considerably more power than the baseline simulation (Figure 10b) at these frequencies. Figure 10c shows the partial spectral profile for model simulation with stormflow; power increases over the baseline run for harmonics 200 and higher. These frequencies represent responses on time scales shorter than nine days, including quickflow time scales. For model simulation with the MSCP (Figure 10d), power increases preferentially for harmonics below 400—time scales of four days or longer. Simulation with both stormflow and the MSCP (Figure 10e), appears to have an additive effect, combining the power increases of simulation with only stormflow and only the MSCP. This power spectrum best matches the magnitudes of the measured runoff record at harmonics above 200, including the highest frequencies representing quickflow time scales. These latter time scales reflect discharge from intial spring snowmelt and stormflow response.

Discussion

Evaluation of Stormflow

The implementation of stormflow alone produces considerable improvement in model representation of catchment discharge. Shallow subsurface stormflow responds to storm events on a shorter time scale than baseflow, and thus allows for a more rapid flushing of waters from the catchment. This stormflow process is critical during dry months. Recharge waters that drain vertically to the deep water table are subject to a very low hydraulic conductivity and thus very low rates of groundwater flow. Consequently, in the absence of stormflow simulation, storm waters are captured by the deep water table and in essence sequestered until wetter conditions prevail, the water table rises, and groundwater flow rates increase. This mechanism explains why models without stormflow work best in wetter conditions when groundwater flow is more responsive. With stormflow activation, however, the catchment is responsive to storm events in both wet and dry conditions. This greater responsiveness results from a partial redirection of vertical recharge waters—instead of first recharging the water table then slowly discharging to river runoff, a portion of the shallow subsurface water moves quickly and directly to streambeds. This effect produces both a redistribution of the timing of discharge and an increase in total runoff generation.

While our conception of stormflow is physically based, the perched water tables, or fully saturated regions, which can develop above the water table and generate stormflow, are not explicitly represented by our modeling approach. Within the land surface, these perched water tables can form in several ways. Heterogeneities in the soil column—different soil types, even sheets of bedrock—produce surfaces of lower permeability upon which water may pool within the soil column. These waters then flow downhill over these lower permeable surfaces. This lateral flow continues only as far as the edge of the substrate; once this edge is reached vertical recharge of the true water table can resume (Noguchi et al., 1999). Because TOPMODEL formulations work from the mean, the simulation of such discontinuities in lateral flow is problematic; TOPMODEL does not allow for pockets of perched bedrock. Without some type of parameterization there is no means of allowing for spatial heterogeneity, i.e. the identification of these perched water table surfaces and their edges.

Another contribution of stormflow to catchment discharge is derived from perched water tables that develop in the soil column by virtue of the decay of hydraulic conductivity with depth. This decay produces a convergence of water with downward flow in the soil column. As waters converge, zones of saturation form. Such zones can develop over entire regions of the watershed, allowing for continuous lateral flow to discharge areas. Our methodology makes no effort to identify such perched water tables explicitly; however, it works in a conceptually similar fashion, using regions of near saturation in the vadose zone to generate stormflow. This approach to modeling stormflow is also consistent with experimental evidence showing that macropores, mesopores, soil pipes, and perched rock within the soil column can direct flow both vertically and laterally (Noguchi et al., 1999; Sidle et al., 2000). While the TOPMODEL approach does not detail such flow pathways, our stormflow conceptualization provides an implicit representation of such lateral flow in the vadose zone.

The convergence of water within the soil column, which can generate stormflow, is analogous to infiltration-excess (Hortonian) runoff production at the surface. These mechanisms are similar in that water accumulates from above faster than it can permeate a soil layer below, and instead must be discharged laterally. Both processes are represented in our model—full saturation of the top layer not only generates stormflow in layer one but also allows for Hortonian overland flow of any continued excess precipitation. Our implementation of stormflow thus smoothly merges with the existing mechanism for generation of infiltration-excess runoff (see Stieglitz et al., 1997) and should be applicable to areas dominated by Hortonian runoff, such as the U.S. Southwest.

Evaluation of the Modified Soil Column Profile (MSCP)

Application of the decrease in both porosity and field capacity with depth in the soil column produces a more responsive water table and thus intensifies groundwater flow discharge. The MSCP modifications impact different aspects of the model simulation than stormflow. Response at the quickflow time scale is not engendered; rather, baseflow becomes more reactive. Instead of a redistribution of the timing of discharge, the magnitude of response increases. Large storm events can be better simulated due to a more responsive water table, and transition recessions—from wet to dry conditions—are improved by this model modification. However, unlike simulation with stormflow, few additional storm events are captured with the MSCP, and early spring wetting from snowmelt is not depicted. In addition, during drier conditions, the model still underrepresents discharge between storm events.

Our sensitivity analyses show that differences in catchment response to the MSCP exist. The steeper catchment of Black Rock with its shallower upslope soils requires a greater rate of decrease in porosity and field capacity to best simulate discharge. This difference in mean profile is warranted in the single column TOPMODEL framework. Because the MSCP is a one-dimensional parameterization of structural variability in upland and lowland regions, physical differences among catchments will require different parameterizations. The MSCP parameterization does not violate TOPMODEL assumptions and thus may be appropriately calibrated.

Better hydrologic simulation might be achieved with a more precise determination of the soil column profiles for porosity and field capacity. The geometric decay functions adopted in this study are admittedly somewhat arbitrary. The form of these functions was chosen to depict a combination of effects—differences in depth to bedrock among upland and lowland areas, as well as bioturbation, and soil compaction—in the absence of detailed experimental evidence. In the future, this profile could be more precisely matched on a catchment-by-catchment basis to analyses of soil samples. These matched functions would need not be monotonic.

Evaluation of the Joint Stormflow/MSCP Application

Combined implementation of both stormflow and the MSCP produces the greatest improvement in model simulation of catchment discharge. In these simulations, model response to storm events during dry conditions better matches measured discharge intensities and durations. Water that would have otherwise been lost to the deep water table is discharged via stormflow, and the MSCP intensifies this response for larger events. The magnitude and timing of spring snowmelt response is also much improved. Stormflow responds quickly to these wetting events, and MSCP permits faster groundwater flow reaction to persistent wetting such as during snowmelt. These two effects amplify one another. Finally, recession curves during the transition from wet to dry conditions are also better represented. The more dynamic water table, produced by the MSCP, effects a more realistic groundwater response during these periods.

Sensitivity analyses also show that different optimum MSCPs exist for the two catchments. As for the model runs with the MSCP alone, this variability reflects differences in catchment structure and relative soil depths among upland and lowland areas.

Simulations with altered soil column layer resolutions demonstrate that the impacts of stormflow and the MSCP are robust within the constraints of the model layering scheme. We do not suggest that the stormflow and MSCP methodologies are entirely resolution independent; however, provided there is sufficient discretization of the soil column, stormflow and groundwater flow discharge simulation appear to be relatively insensitive to changes in layer thicknesses.

Summary

This study has presented two new modeling strategies: a methodology for representing subsurface stormflow; and a modified soil column profile (MSCP) that produces a more responsive water table. Both stormflow and the MSCP are adopted in a manner consistent with TOPMODEL assumptions. The stormflow methodology allows for discharge from regions of near saturation that develop in the vadose zone during storm events. Determination of the stormflow contribution to discharge is made using the equation for groundwater flow, and is a function of catchment topography and local hydraulic conductivity at the depth of such regions of near saturation. Stormflow simulation produces discharge at quickflow time scales.

The MSCP represents the different soil column profiles of porosity and field capacity in upland and lowland regions of a watershed in the single column TOPMODEL framework. This parameterization of physical structure creates a more dynamic water table and more responsive groundwater flow. In applying the MSCP, large storm events are better captured and transition recessions from wet to dry conditions are better simulated. Steeper catchments with shallower upland soils appear to require a greater reduction of porosity and field capacity to better simulate large storm event discharge.

The new modeling strategies have been applied to two experimental watershed in the northeastern United States. Used jointly, stormflow and the MSCP provide a more accurate representation of the time scales of catchment response to storm events and a more complete depiction of hydrologic activity. Storm events large and small are better simulated, and some of the biases previously evident in TOPMODEL simulations are reduced.

Acknowledgements

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Figure Captions

Figure 1. Schematic of the hydrology model. The model couples the analytic form of TOPMODEL equations within a discretized column framework. From an update of the mean water table depth, TOPMODEL equations and DEM data are used to generate groundwater flow and the saturated fraction of the watershed.

Figure 2. Schematic of the stormflow and MSCP modifications. a) Baseline model soil column with a fixed level of porosity and field capacity with depth. Porosity and field capacity are represented jointly by the open areas of the column; solid soil is represented by the hatched brick filling the left of the column; in this version vertical infiltration recharges the water table raising groundwater flow. b) Model with stormflow—gravity directs a component of the recharge laterally generating stormflow; c) Model with stormflow and the MSCP—as for b) but with reduced porosity and field capacity with depth (represented by the expanding hatched brick).

Figure 3. Calculation of stormflow. a) Calculation of stormflow in layer i begins if saturation exceeds 70% of field capacity. b) Water in excess of 70% saturation is redistributed in a fully saturated band from the base of layer i upward; all layers below

Figure 1

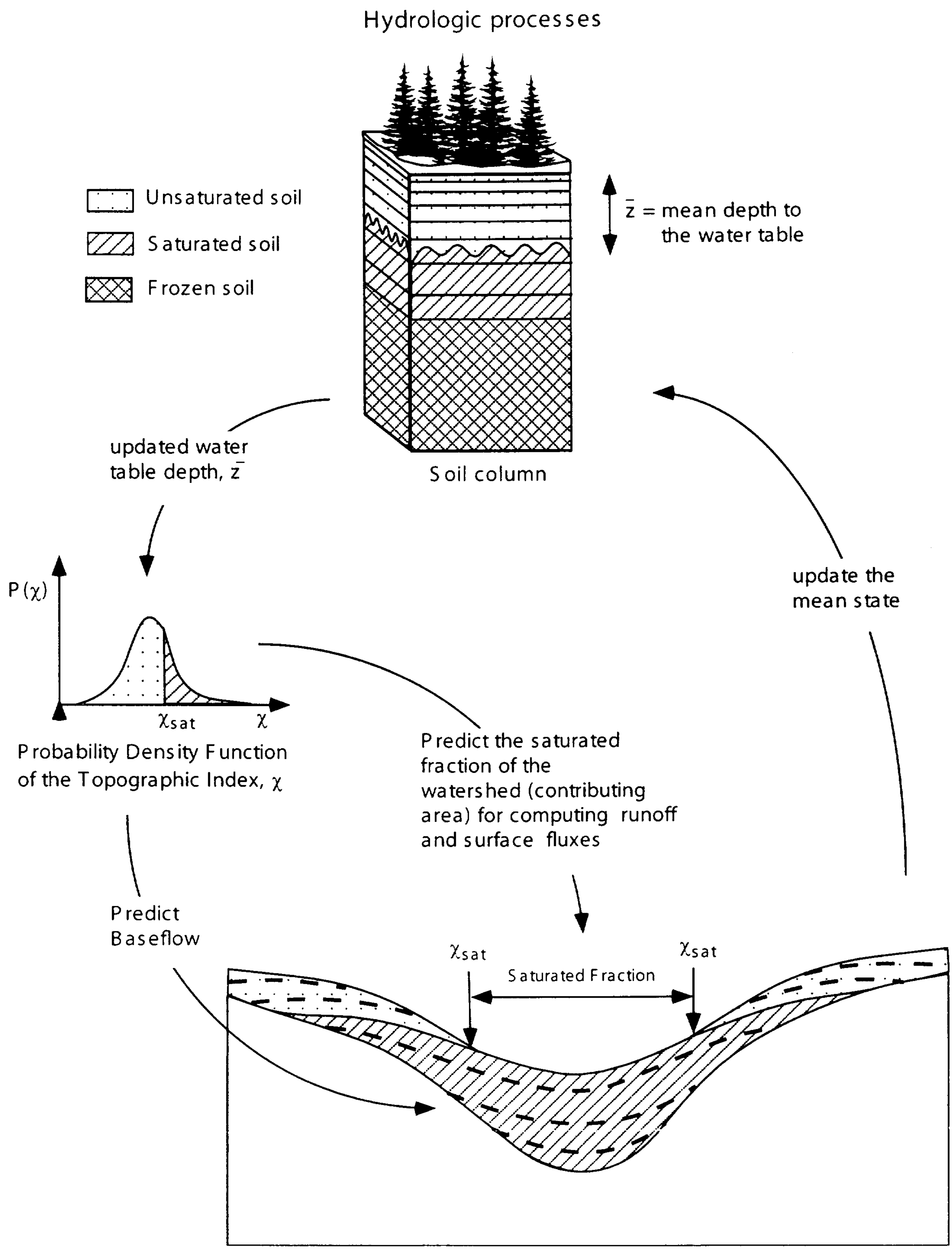

Hydrologic processes
Unsaturated soil
Saturated soil
Frozen soil
$\bar{z}$ = mean depth to the water table
Soil column
updated water table depth, $\bar{z}$
update the mean state
P(χ)
χ_{sat}
χ
Probability Density Function of the Topographic Index, χ
Predict the saturated fraction of the watershed (contributing area) for computing runoff and surface fluxes
Predict Baseflow
χ_{sat}
χ_{sat}
Saturated Fraction

Figure 2

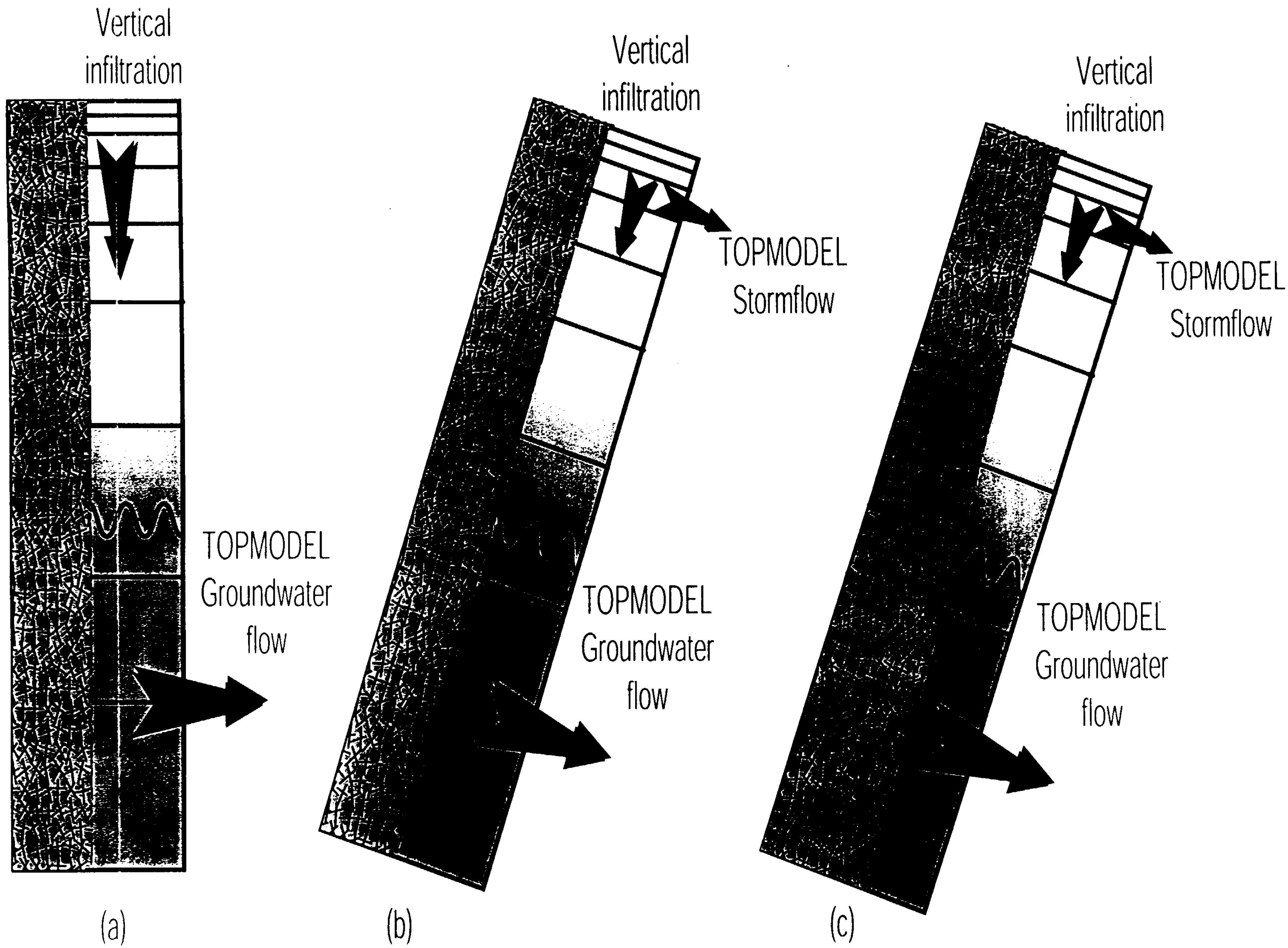
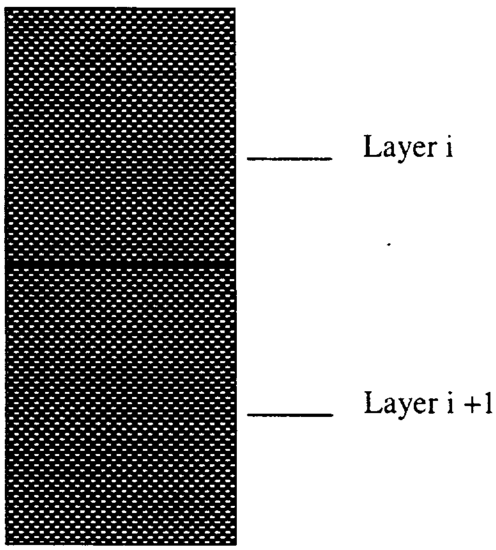
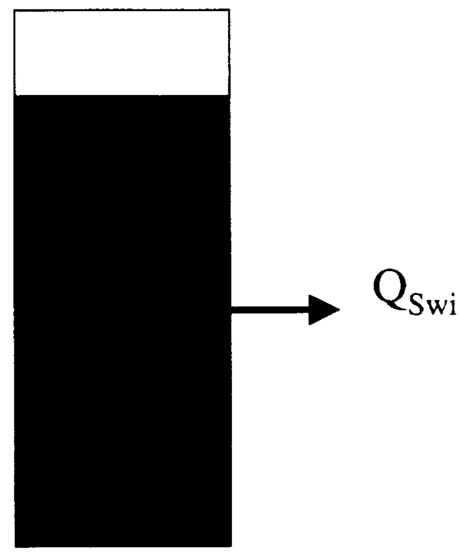

Vertical infiltration
TOPMODEL Groundwater flow
(a)
Vertical infiltration
TOPMODEL Stormflow
TOPMODEL Groundwater flow
(b)
Vertical infiltration
TOPMODEL Stormflow
TOPMODEL Groundwater flow
(c)

Figure 3 Calculation of Stormflow (Q_S)

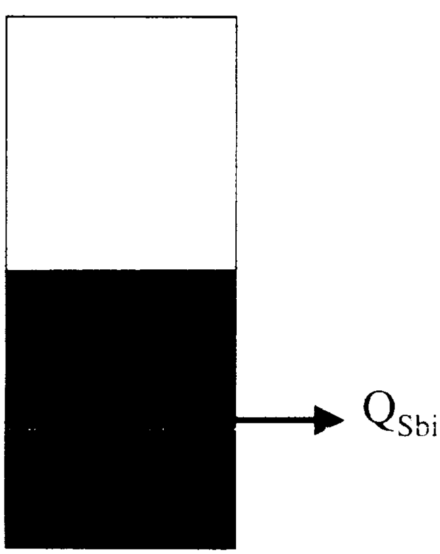
a) Layers in the Vadose Zone



b) Step 1



c) Step 2



d) Step 3:

$Q_{Si} = Q_{Swi} - Q_{Sbi}$

Repeat for Each Layer in the Vadose Zone and Total $Q_S = \Sigma Q_{Si}$

Total Runoff $Q_R = Q_B + Q_S$

1971 Sleepers River Catchment Simulations

Figure 4a
Baseline Model Hydrograph

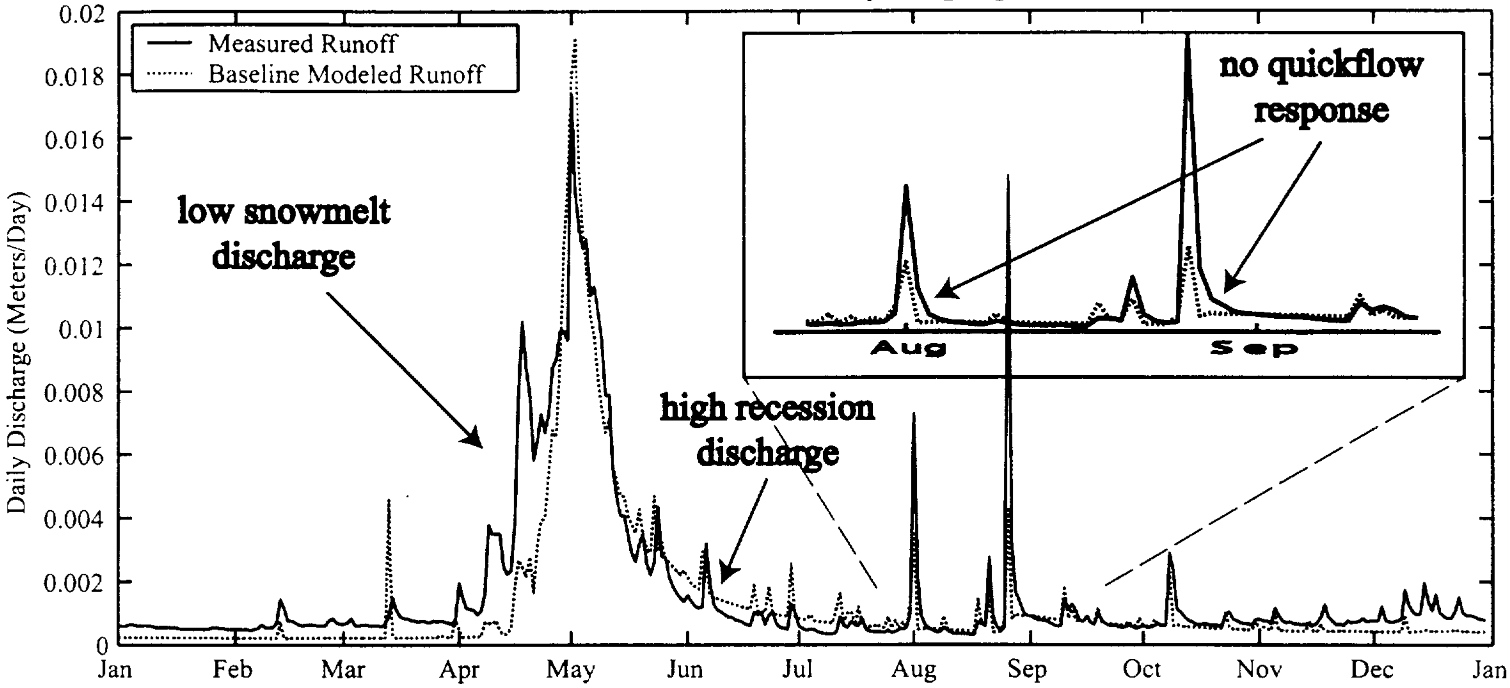
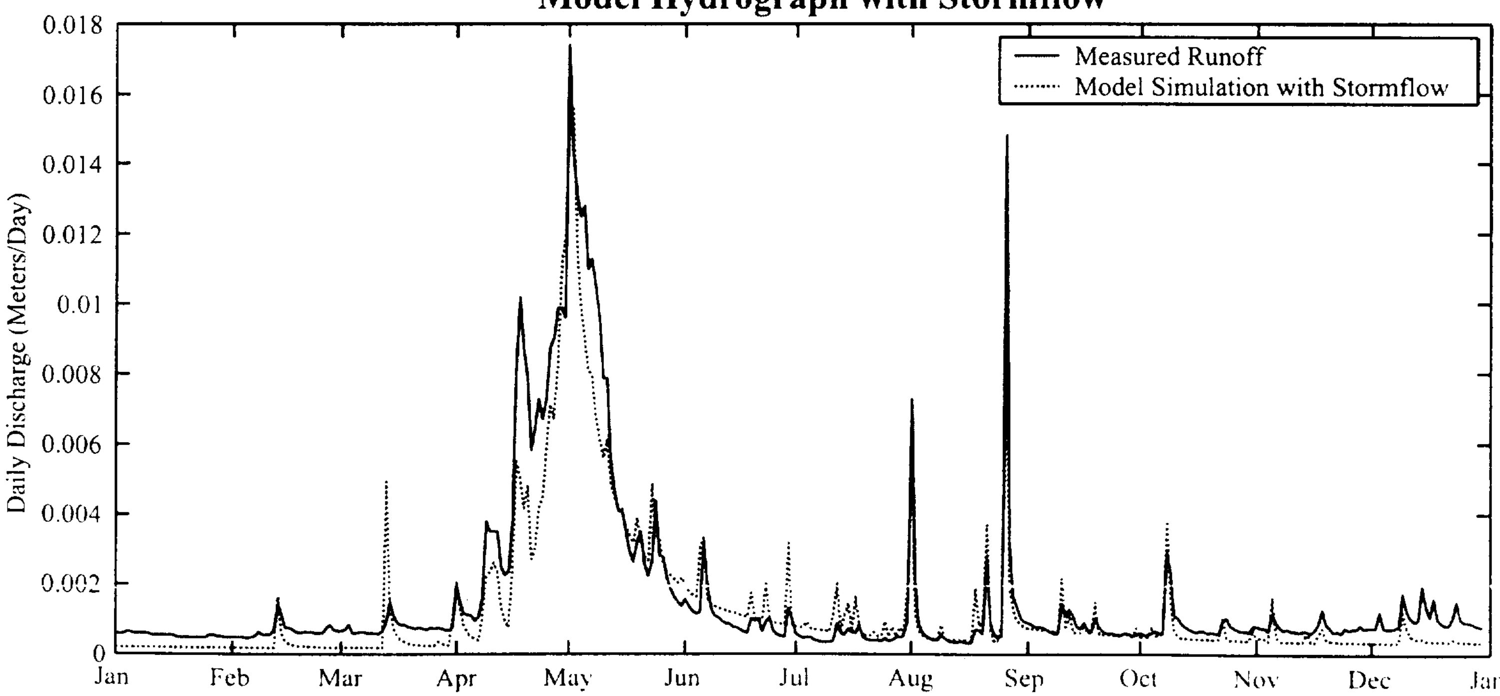


Figure 4b
Model Hydrograph with Stormflow



1971 Sleepers River Catchment Simulations

Figure 4c
Model Hydrograph with the MSCP

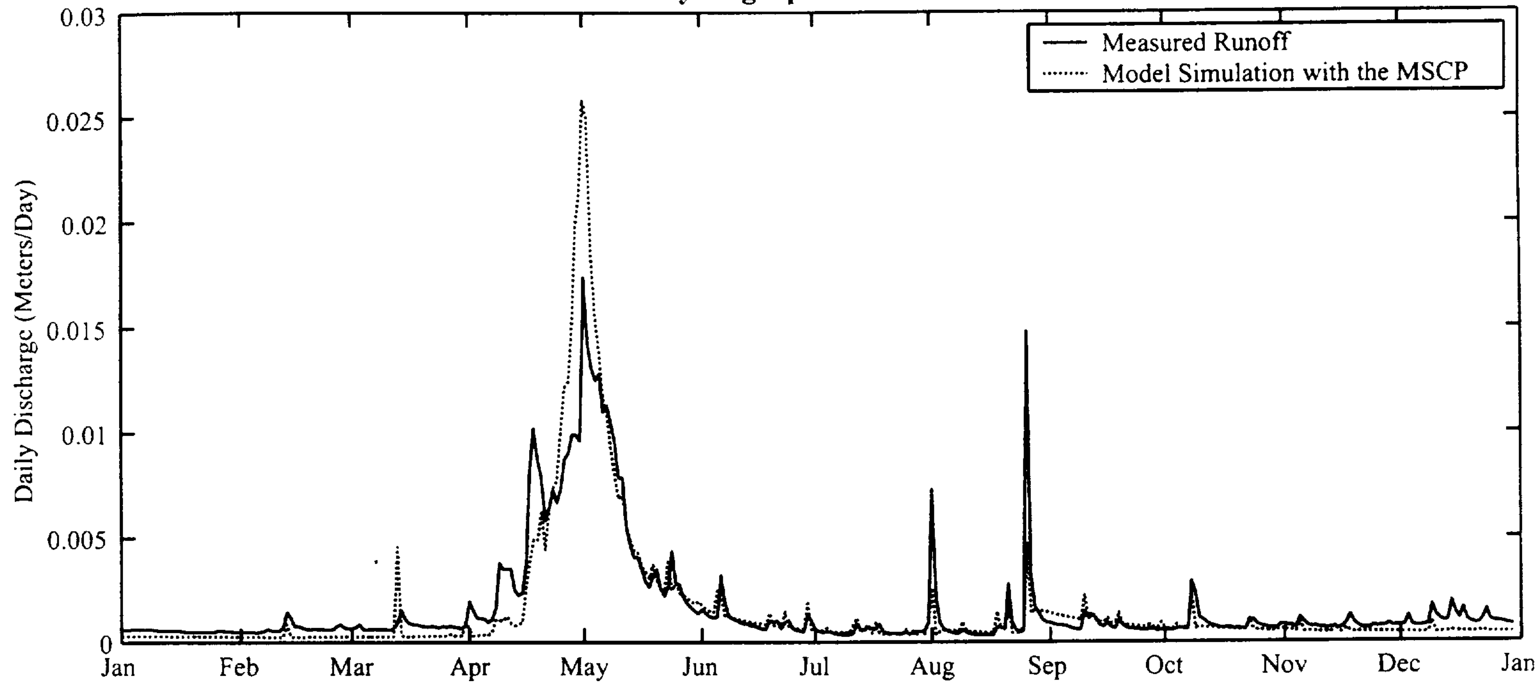
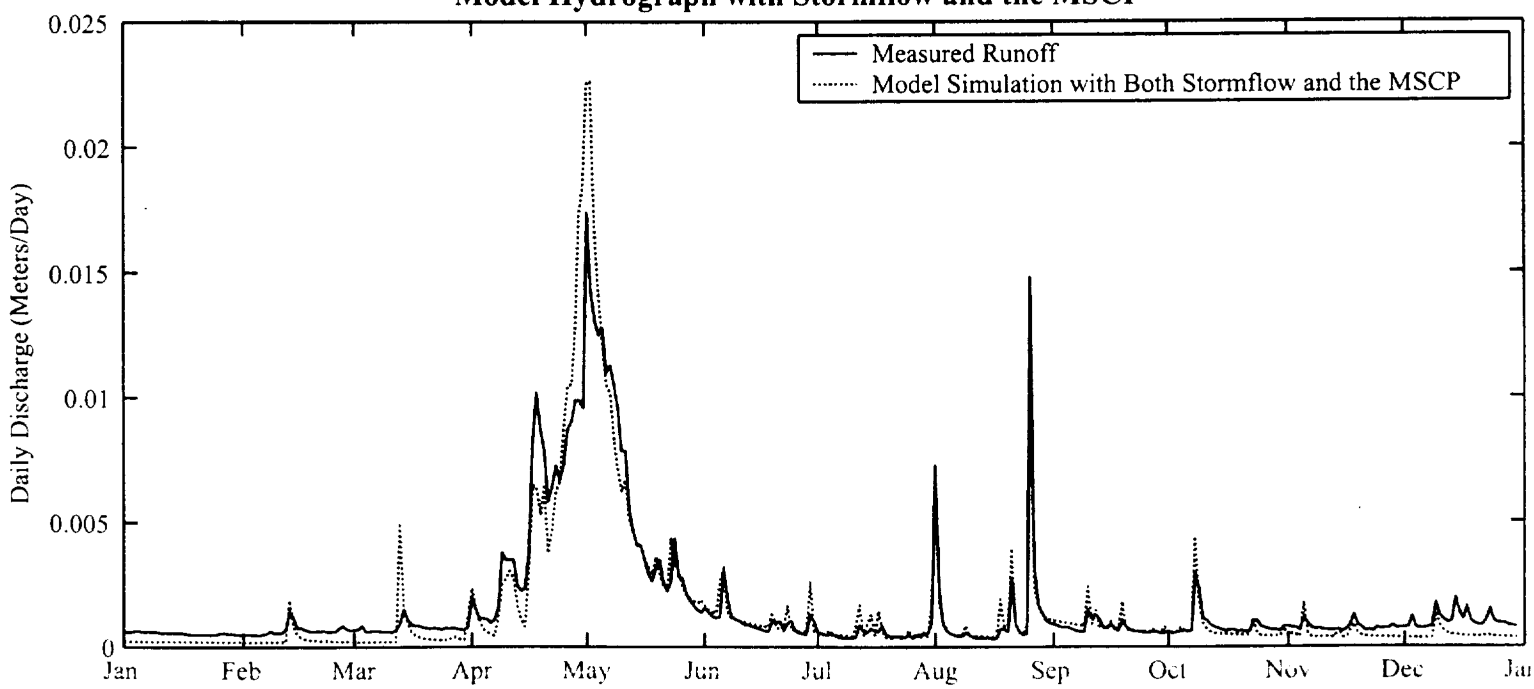


Figure 4d
Model Hydrograph with Stormflow and the MSCP



1973 Sleepers River Catchment Simulations

Figure 5a

Baseline Model Hydrograph

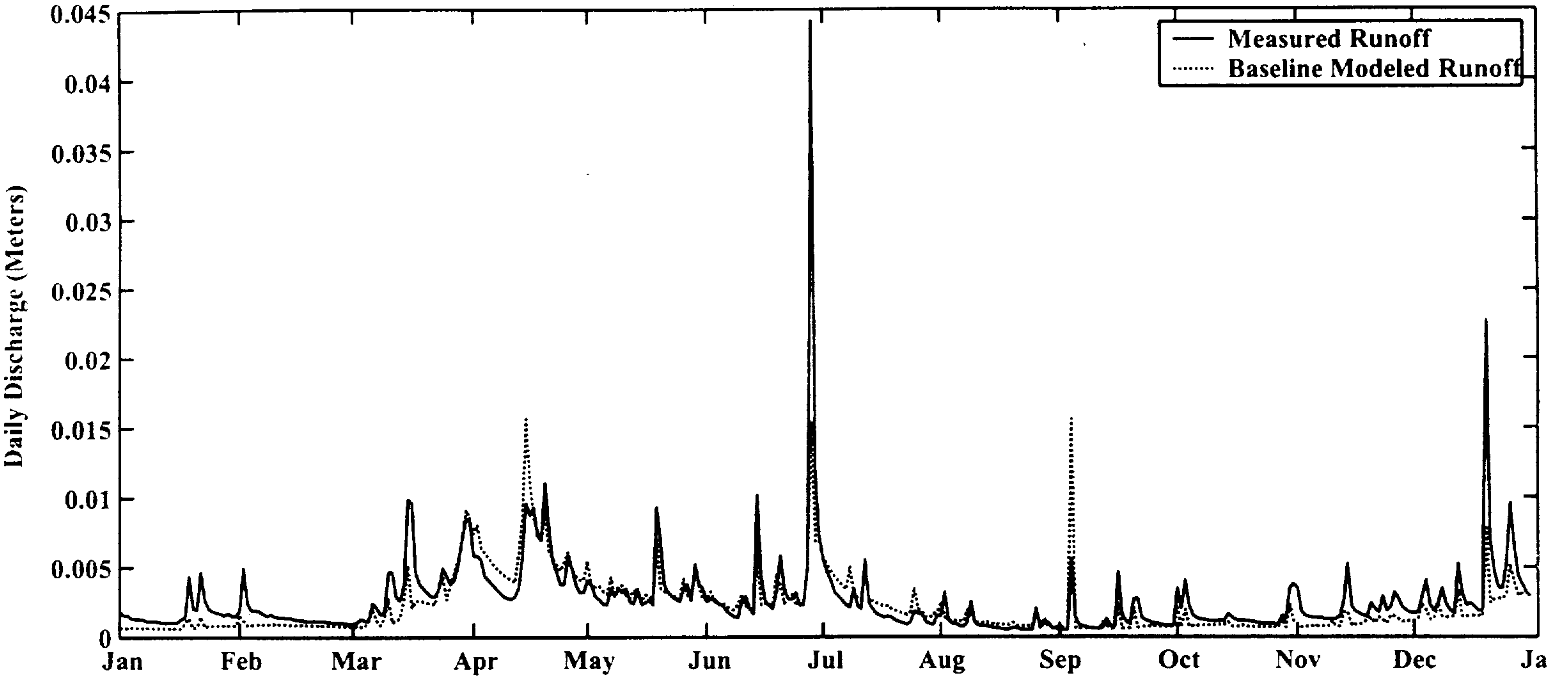
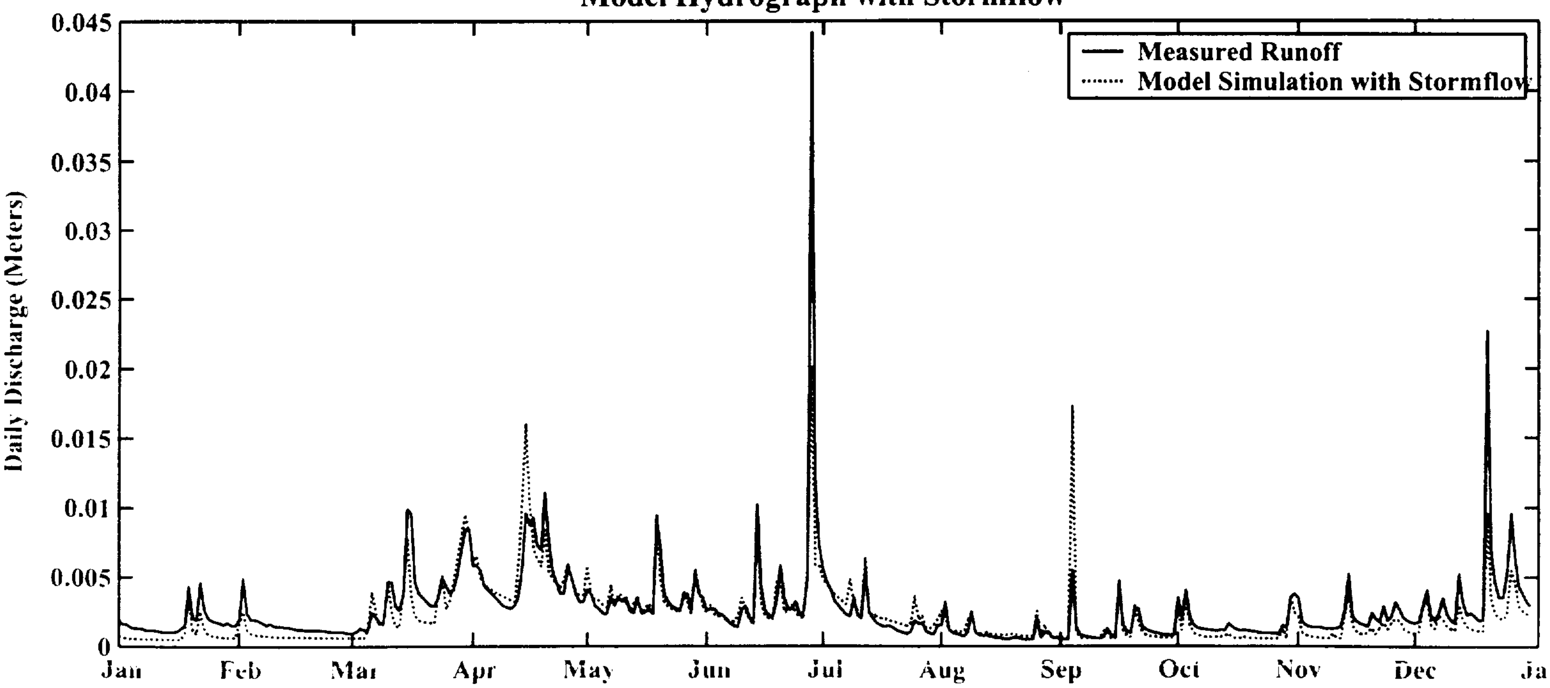


Figure 5b

Model Hydrograph with Stormflow



1973 Sleepers River Catchment Simulations

Figure 5c
Model Hydrograph with the MSCP

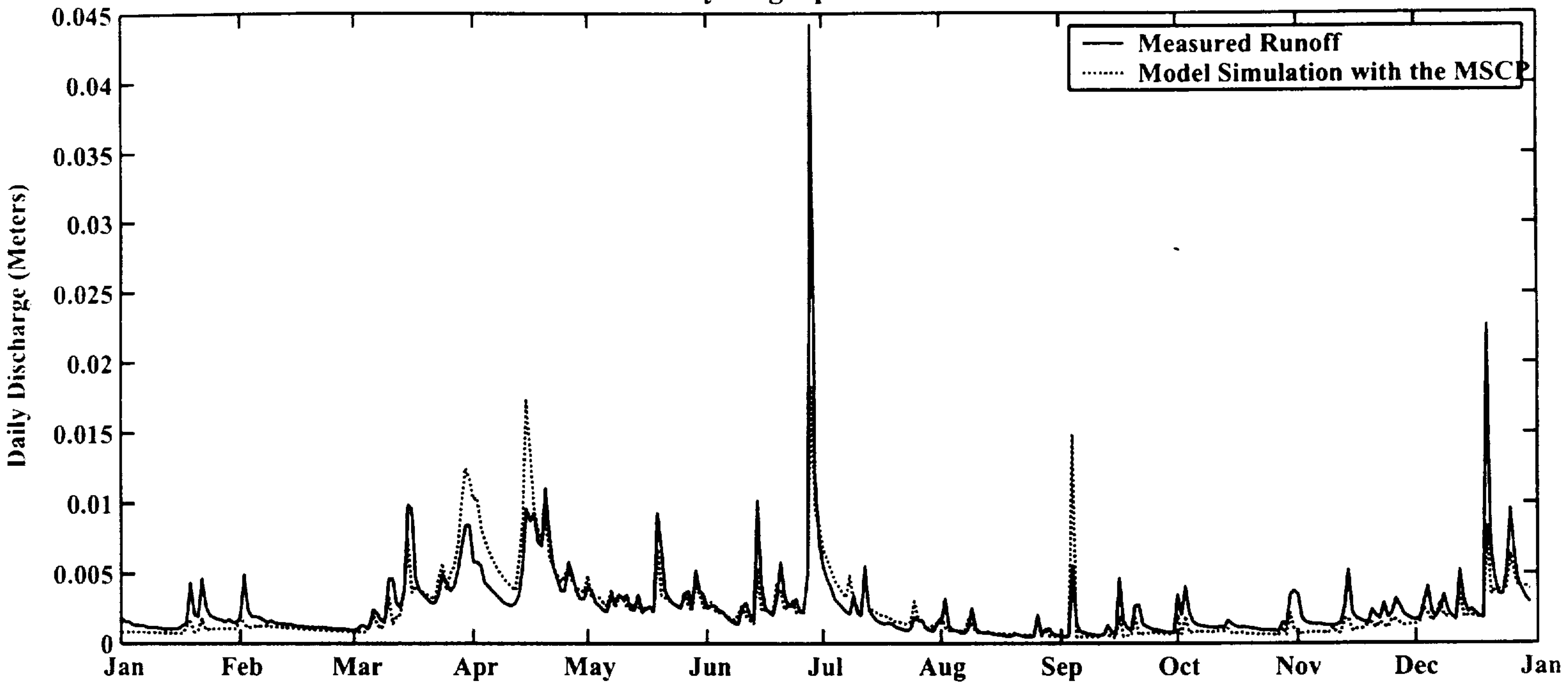
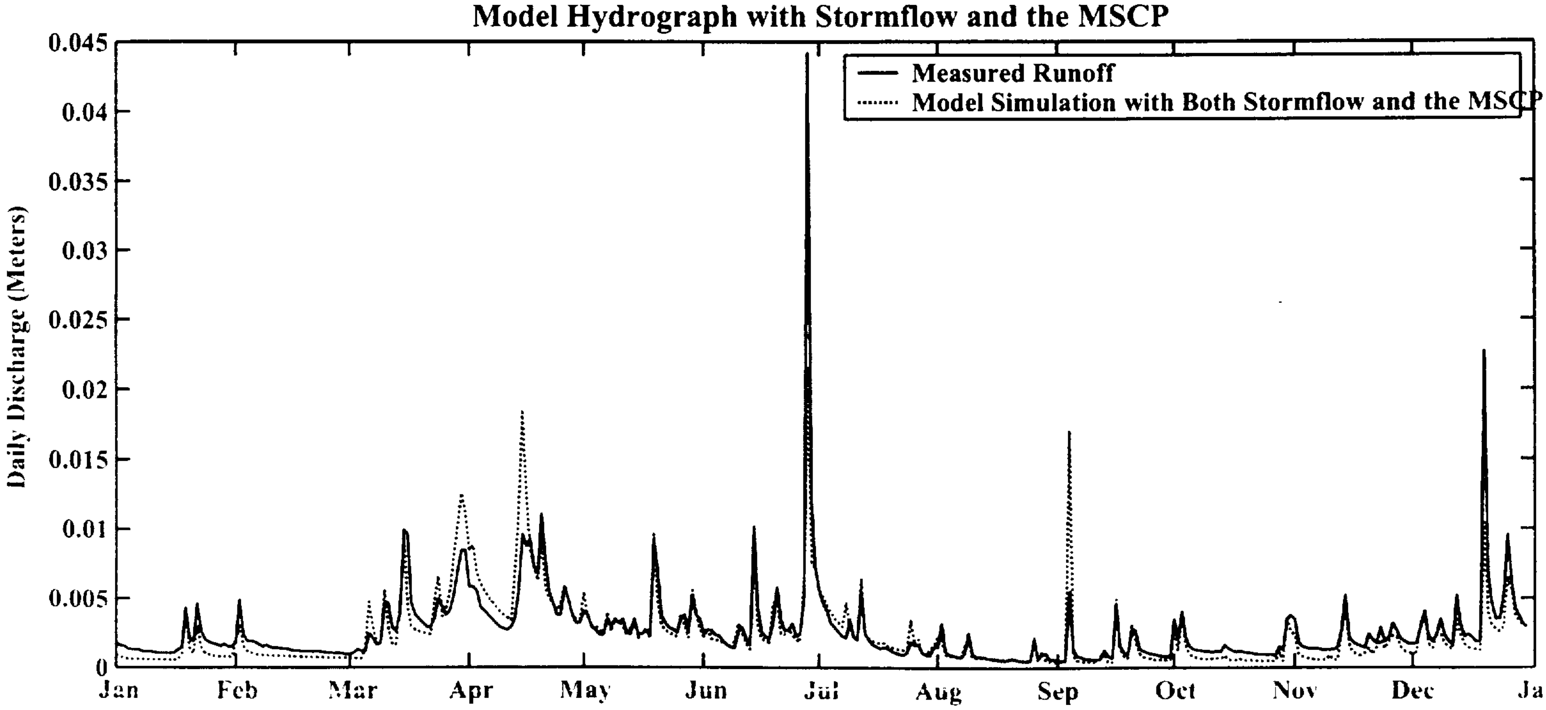


Figure 5d
Model Hydrograph with Stormflow and the MSCP



1999 Black Rock Simulations
Figure 6a
Baseline Model Hydrograph

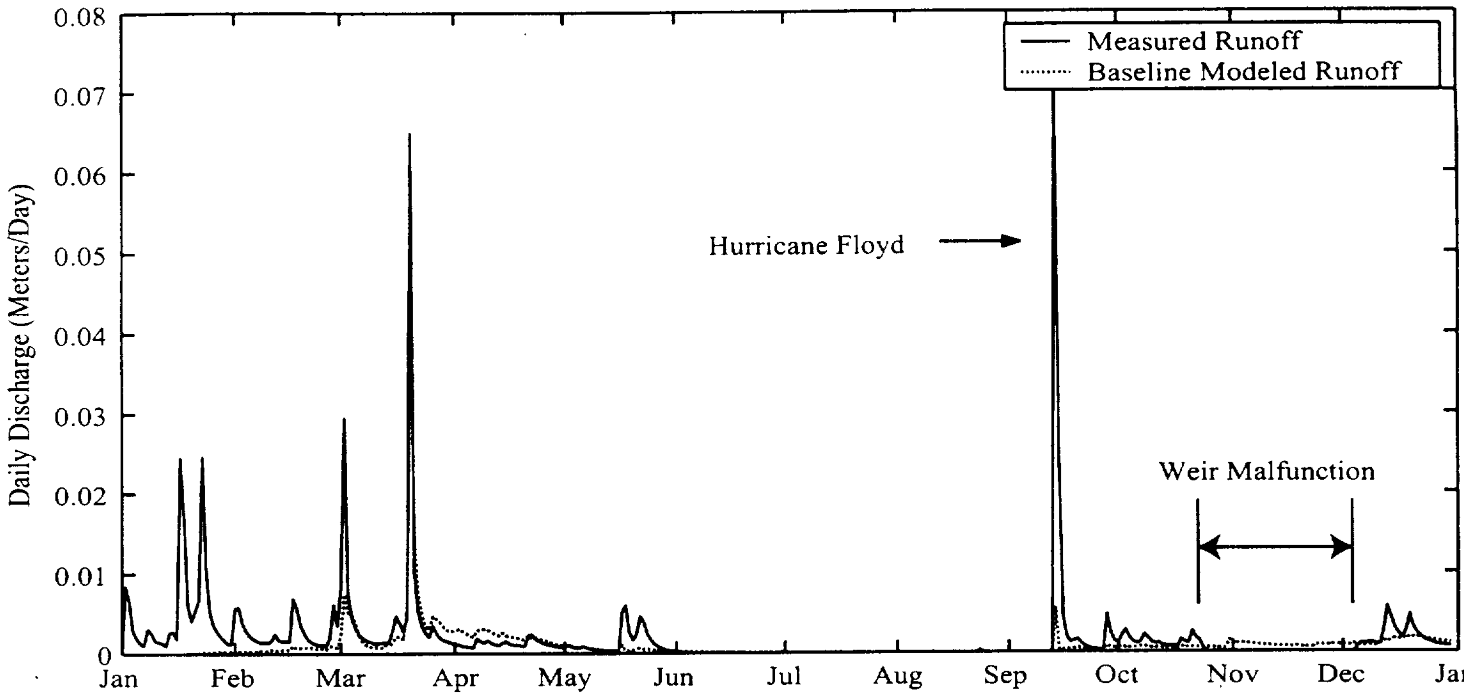
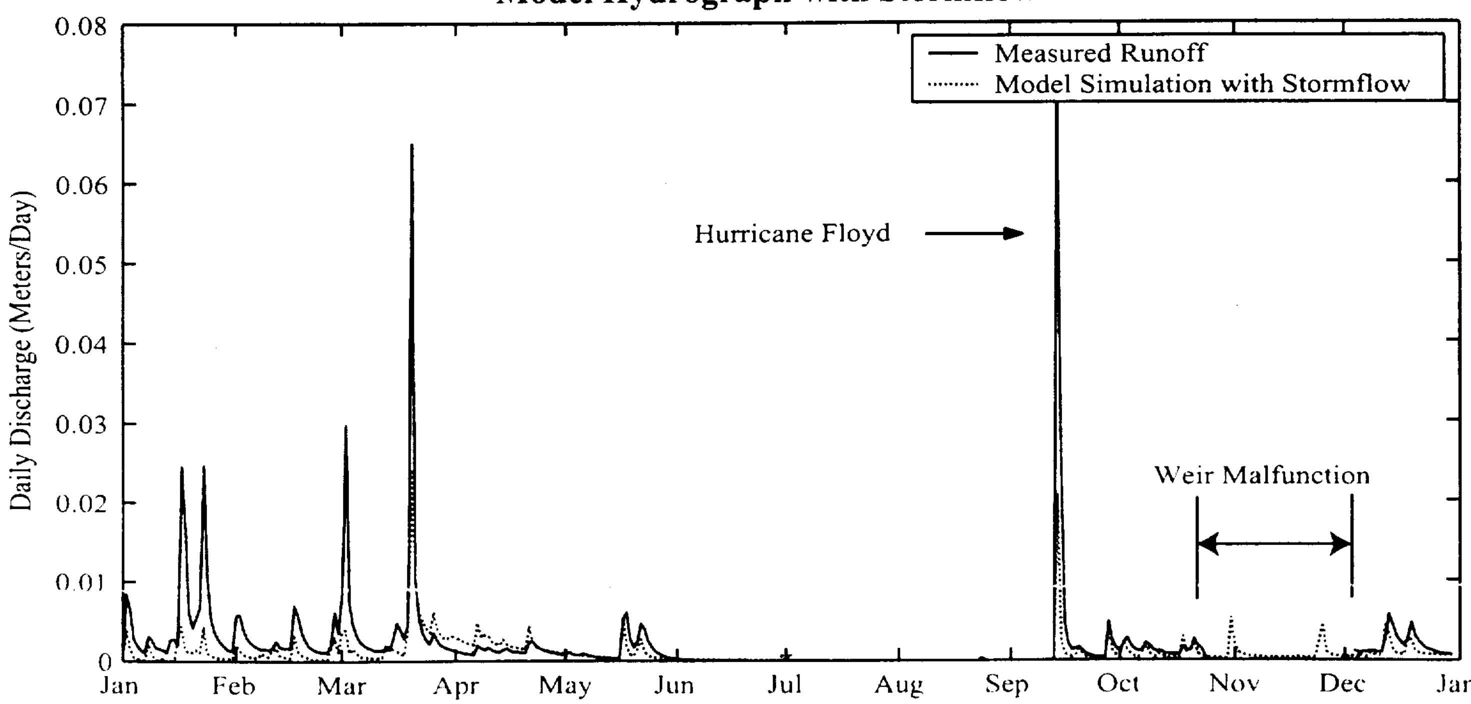


Figure 6b
Model Hydrograph with Stormflow



1999 Black Rock Simulations

Figure 6c
Model Hydrograph with the MSCP

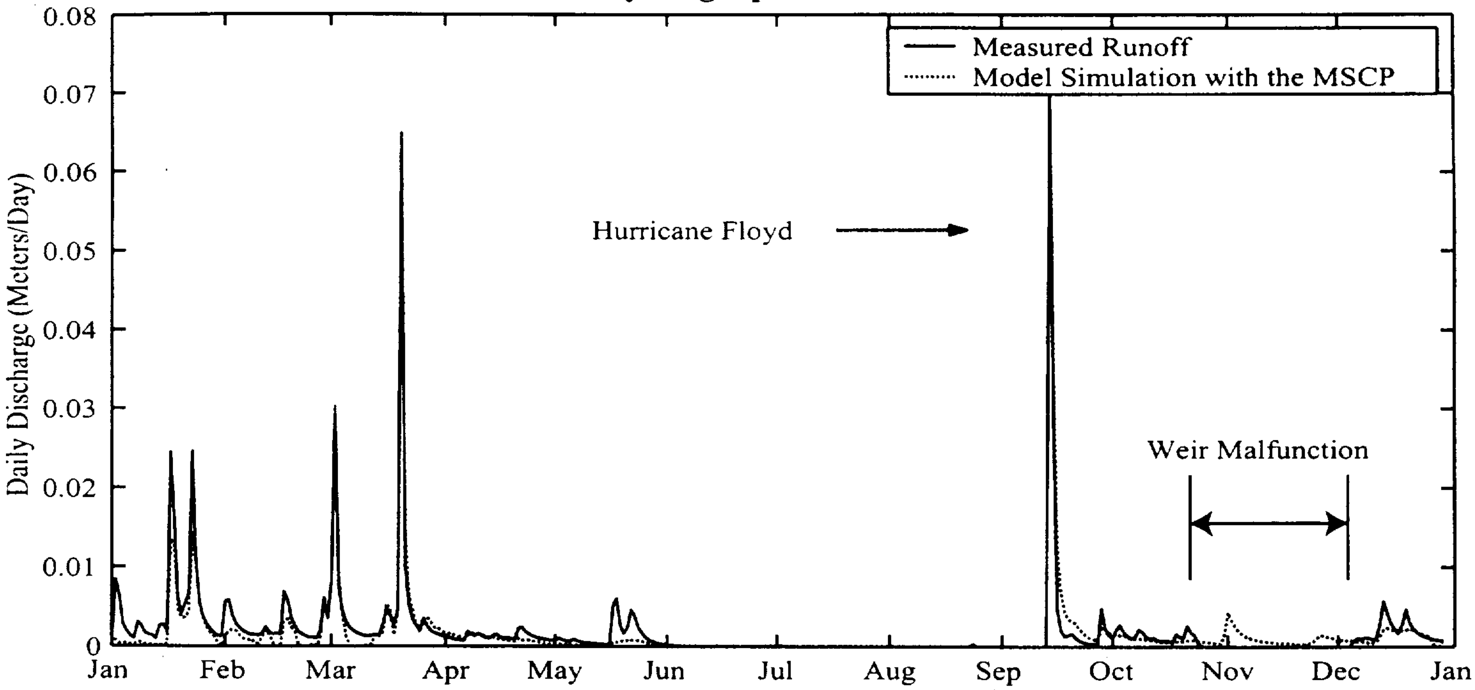
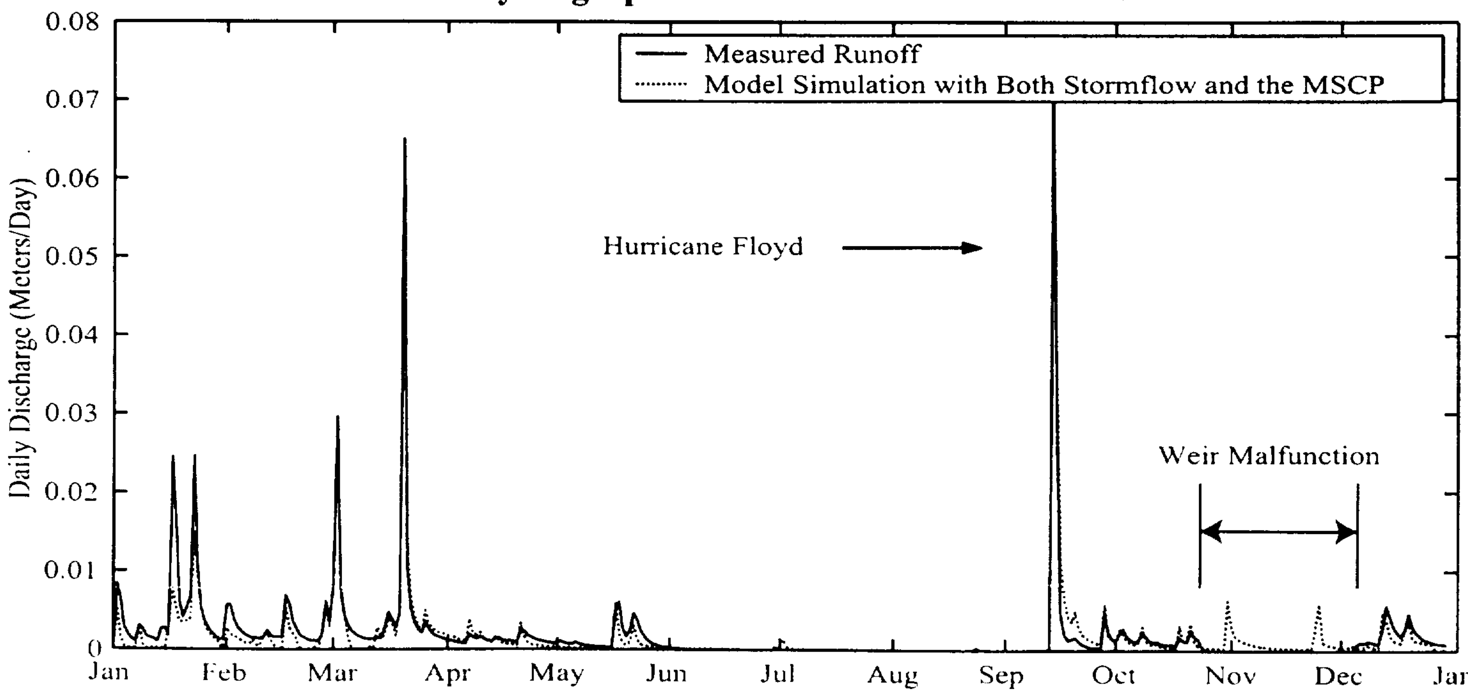


Figure 6d
Model Hydrograph with Stormflow and the MSCP



2000 Black Rock Simulations
Figure 7a
Baseline Model Hydrograph

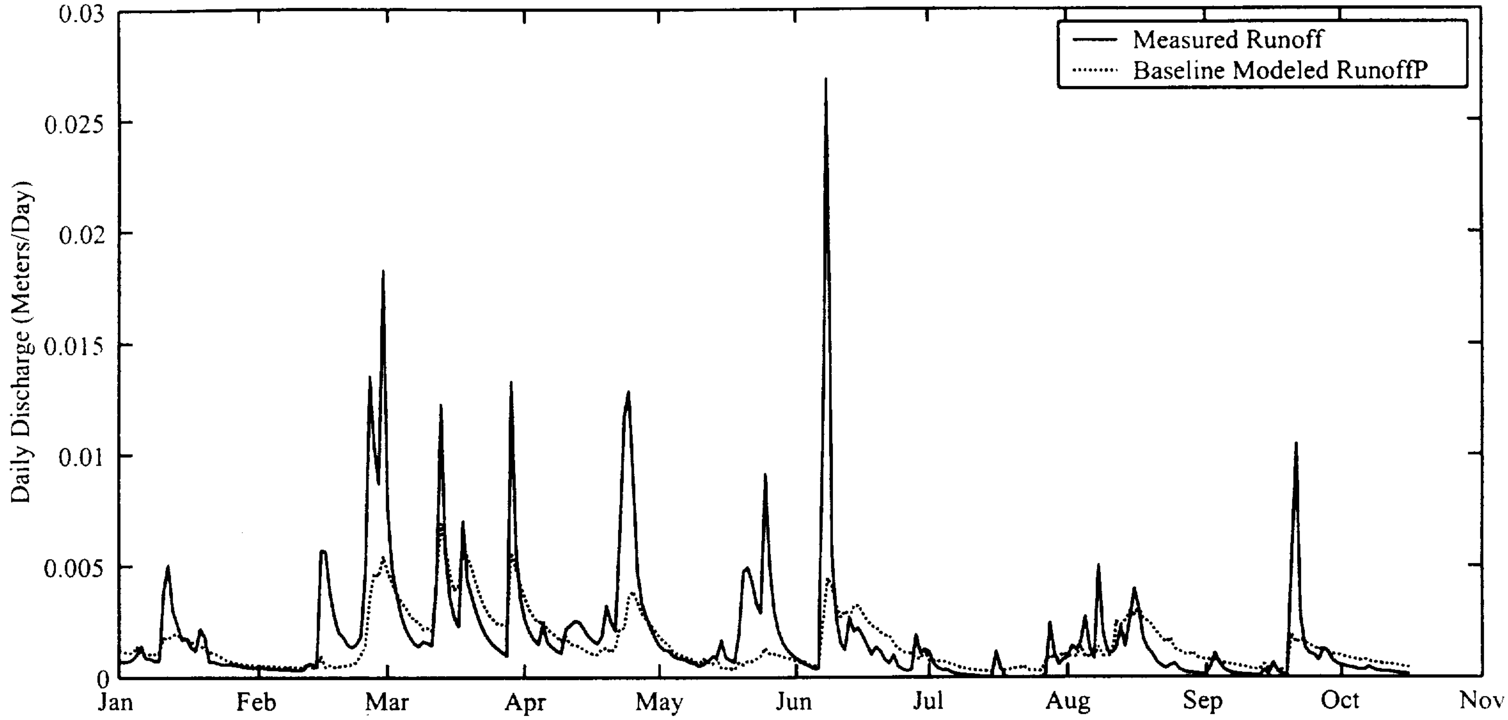
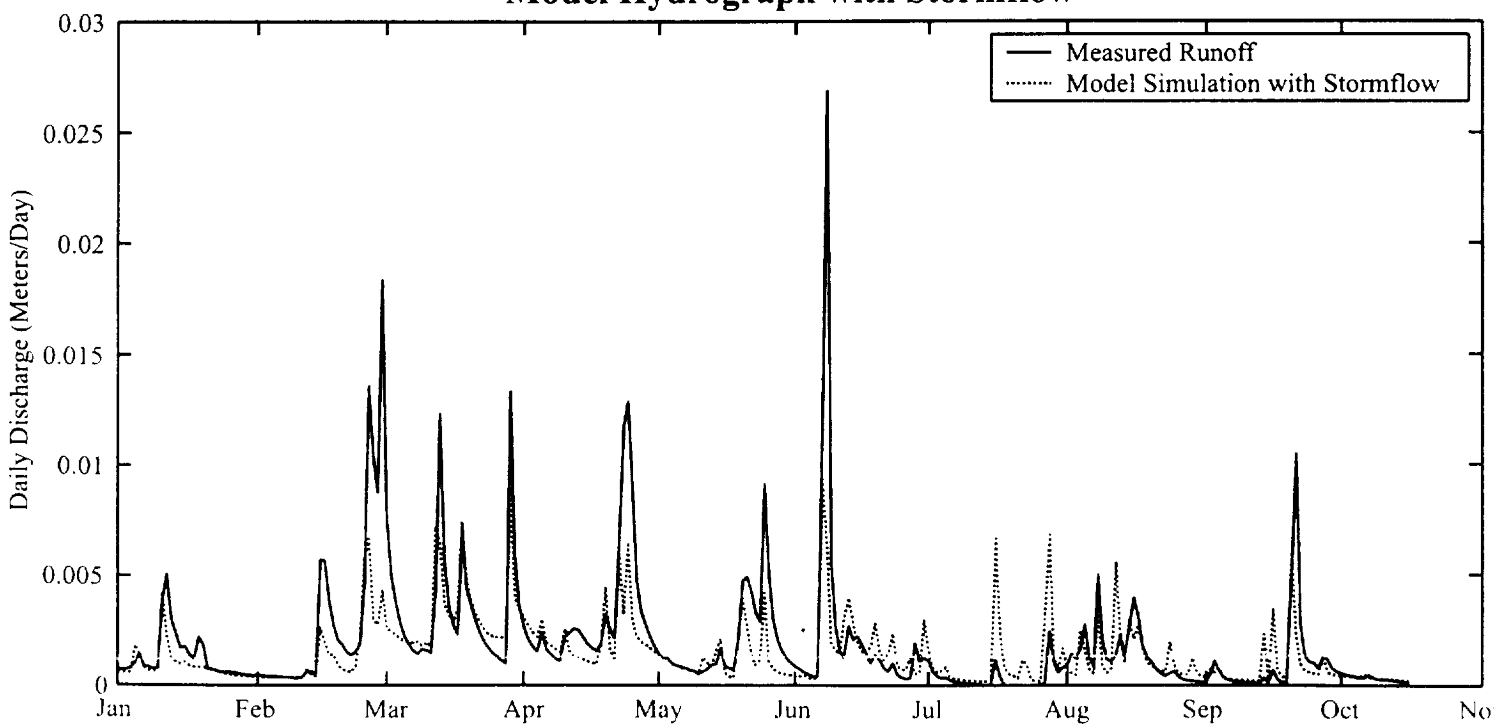


Figure 7b
Model Hydrograph with Stormflow



2000 Black Rock Simulations

Figure 7c
Model Hydrograph with the MSCP

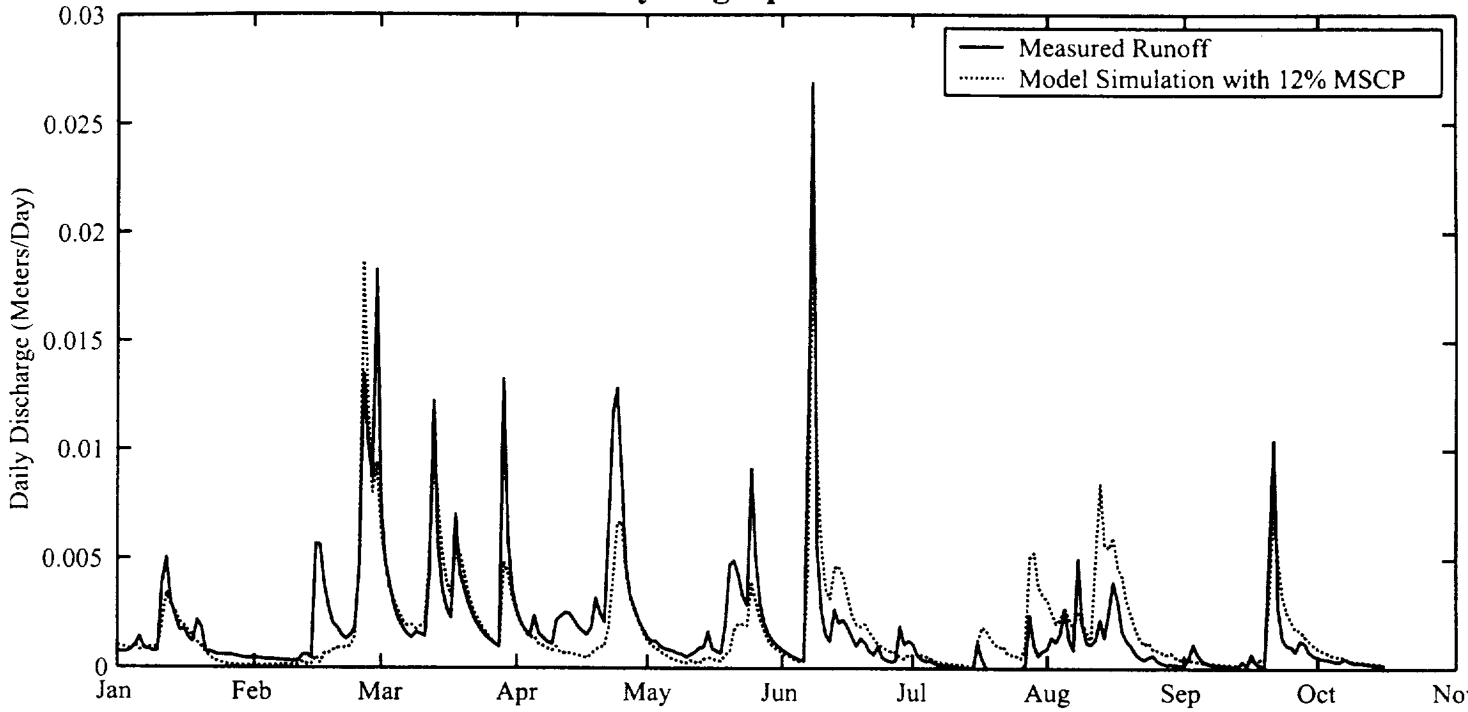
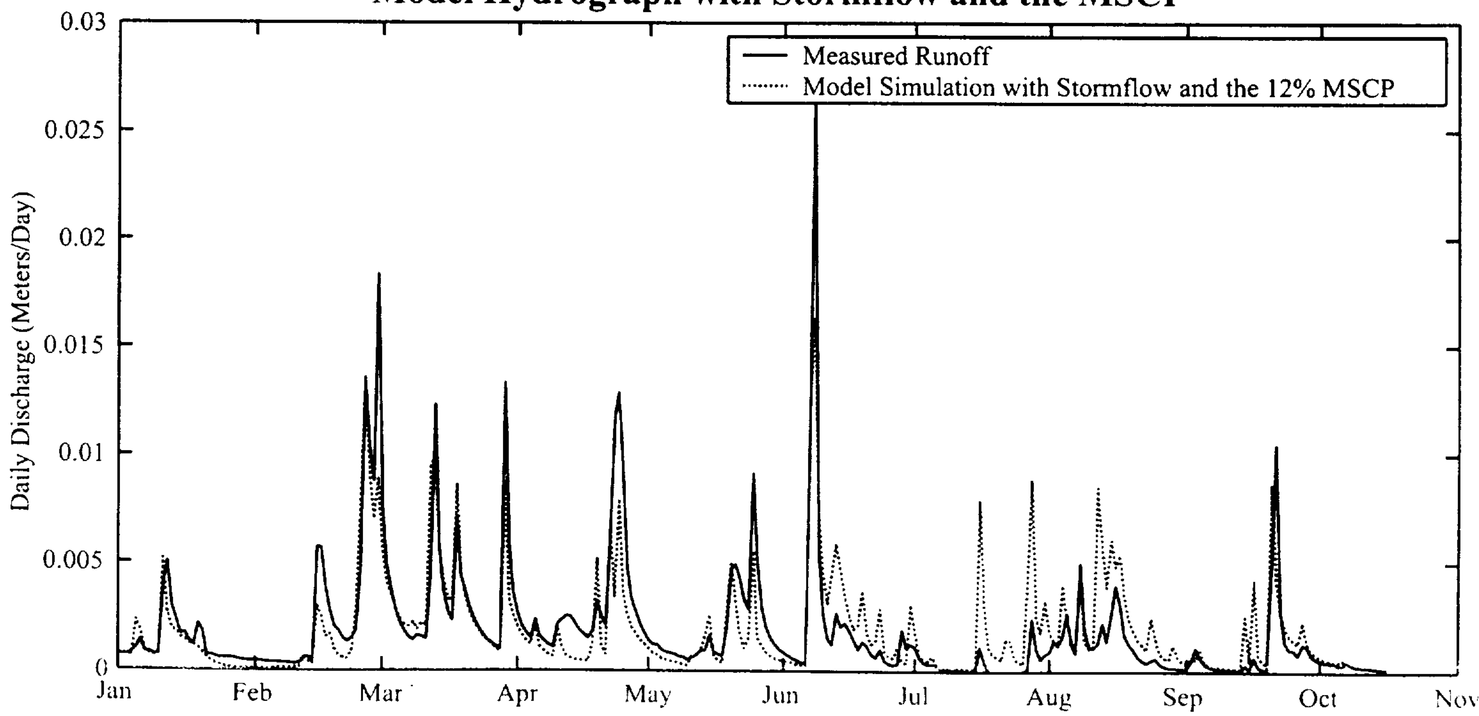


Figure 7d
Model Hydrograph with Stormflow and the MSCP



Sleepers River Catchment Simulations - Mean Water Table Depth

Figure 8a

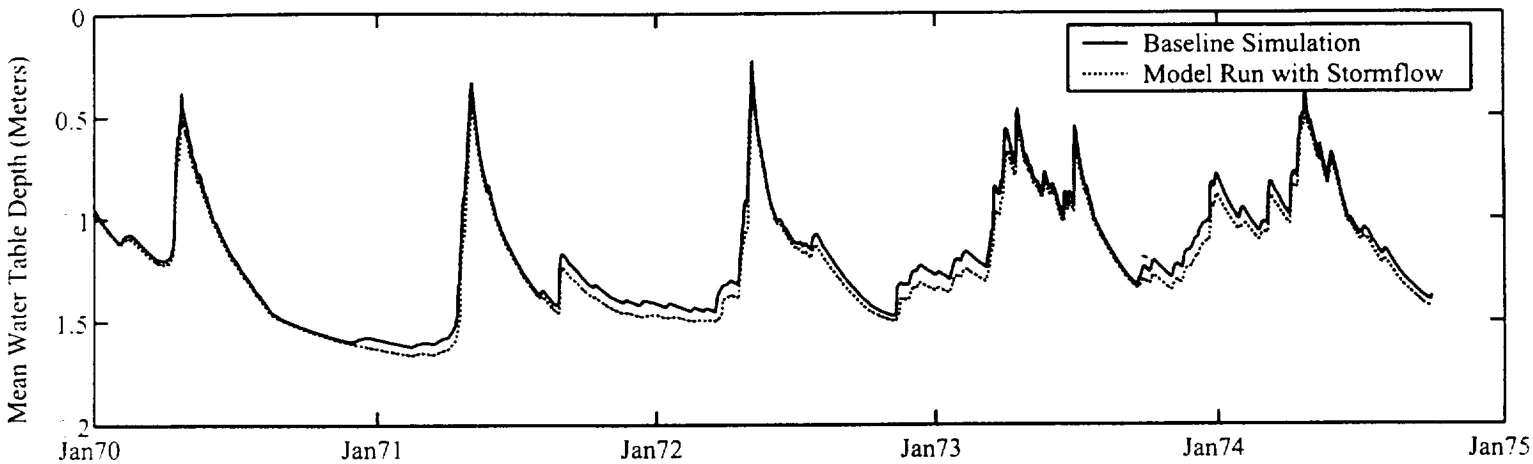


Figure 8b

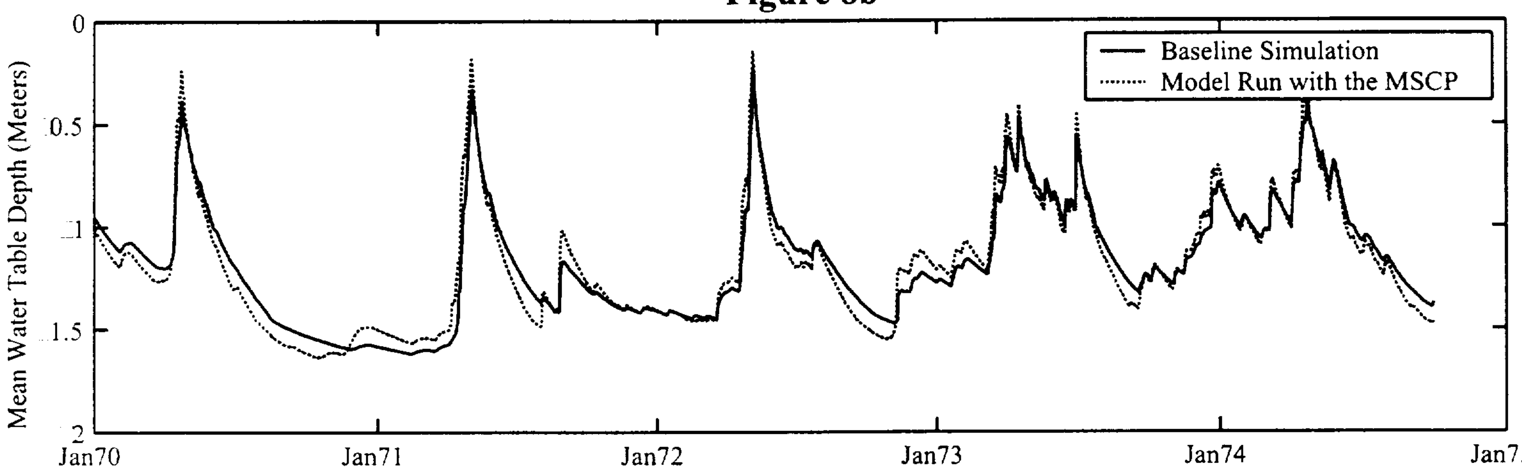


Figure 8c

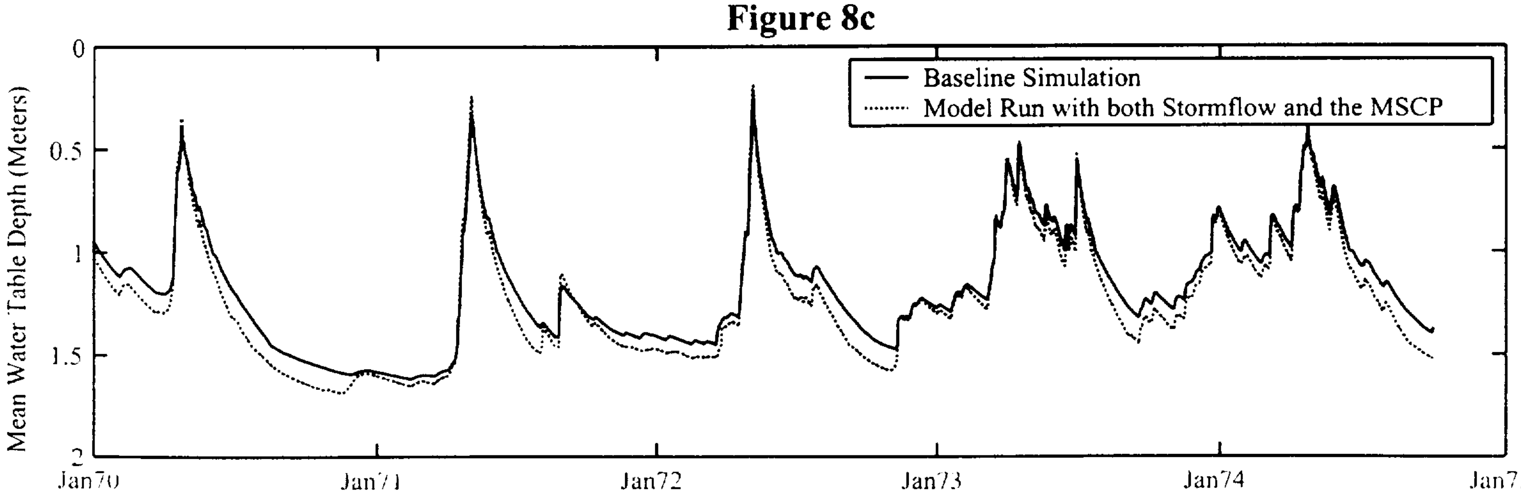


Figure 9a
Sleepers River Measure Runoff

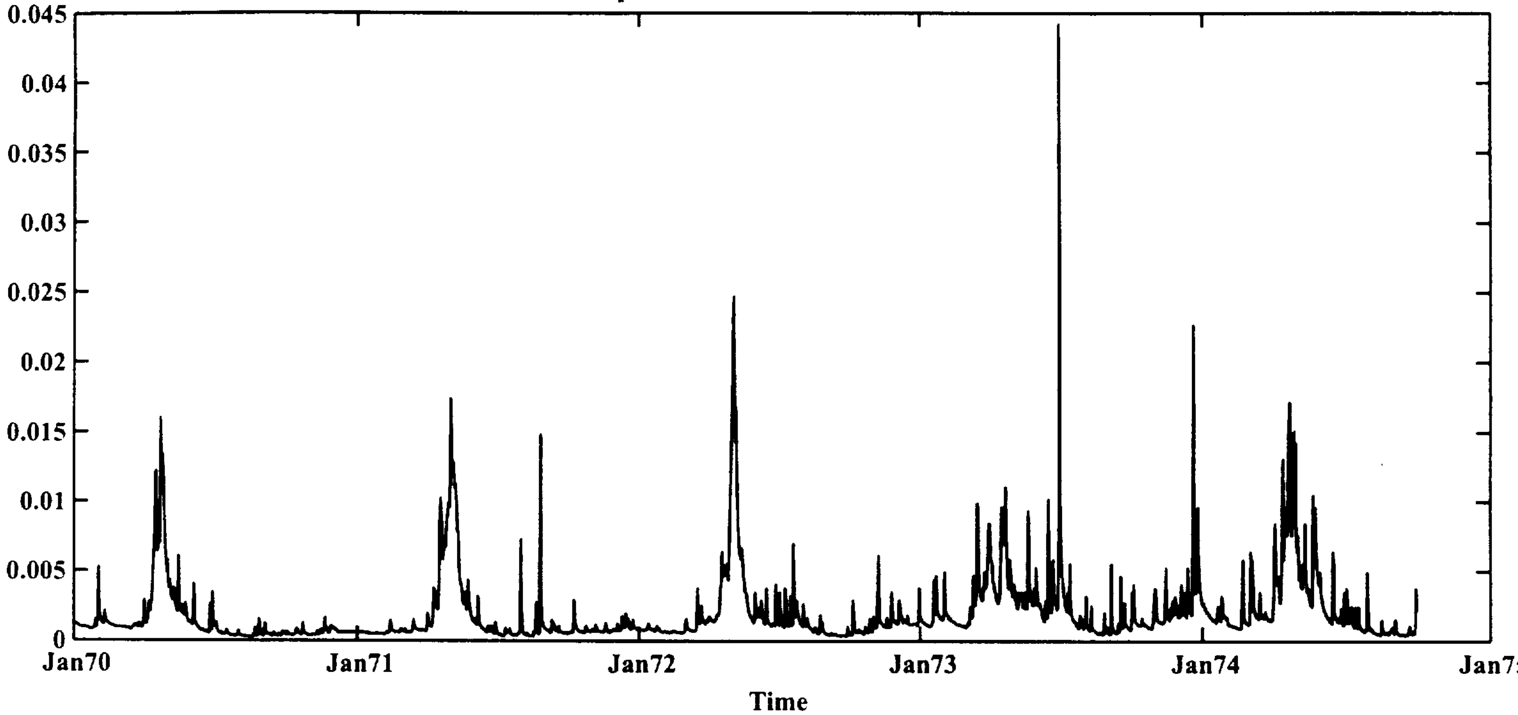


Figure 9b
Spectral Analysis (Full Spectrum)

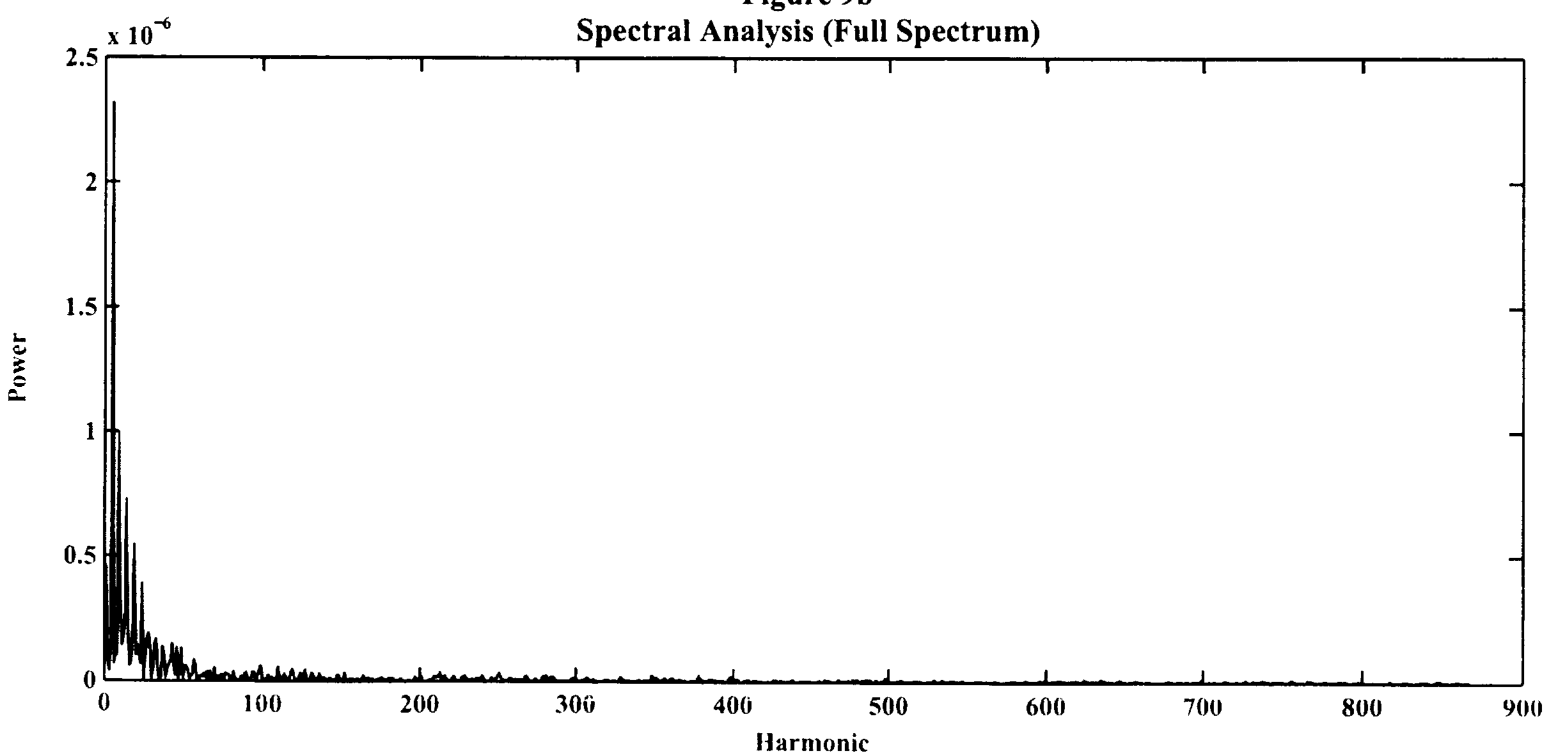


Figure 10a
Spectral Analysis of Sleepers River Hydrographs
Measured Runoff (Partial Spectrum)

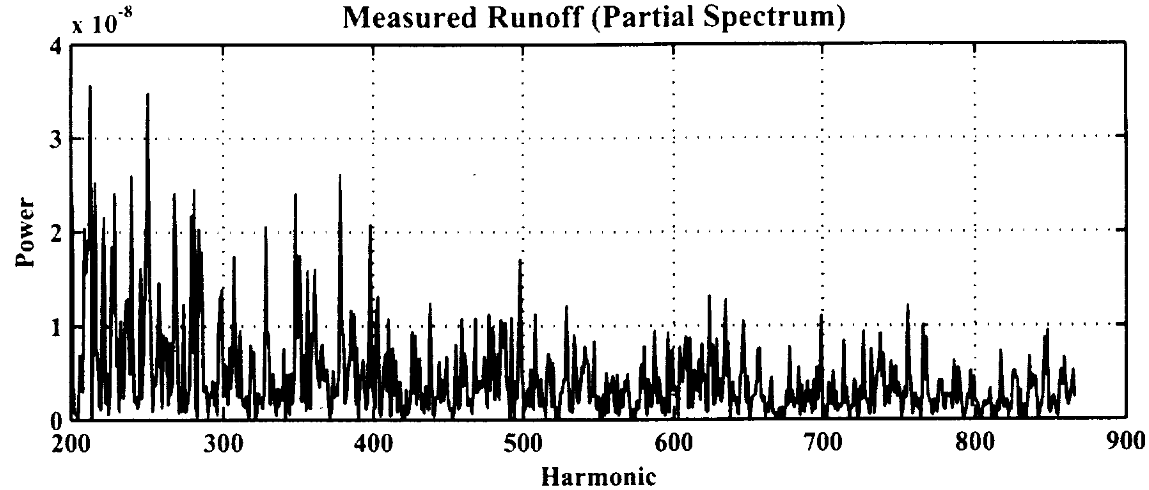


Figure 10b - Baseline Model Simulation

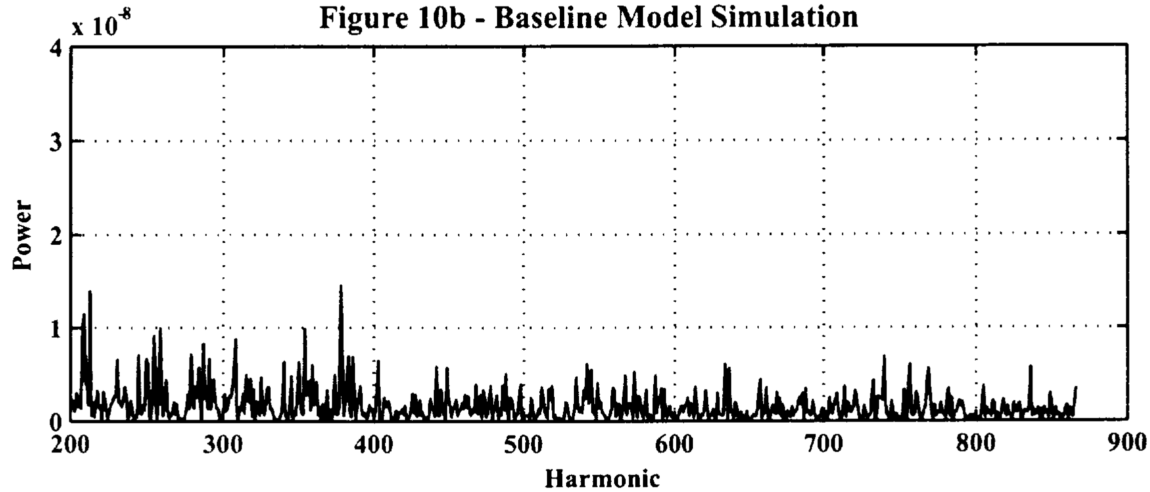
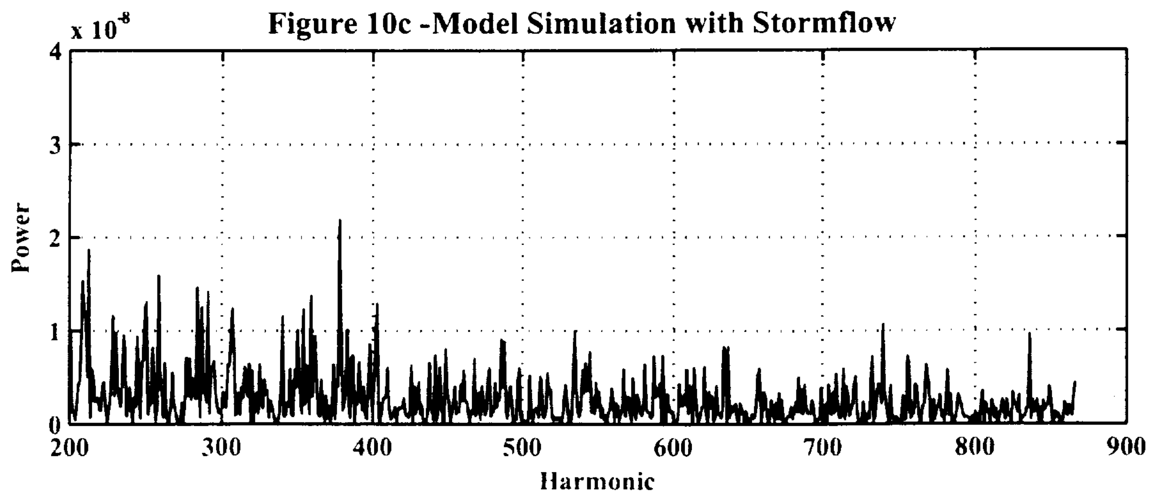
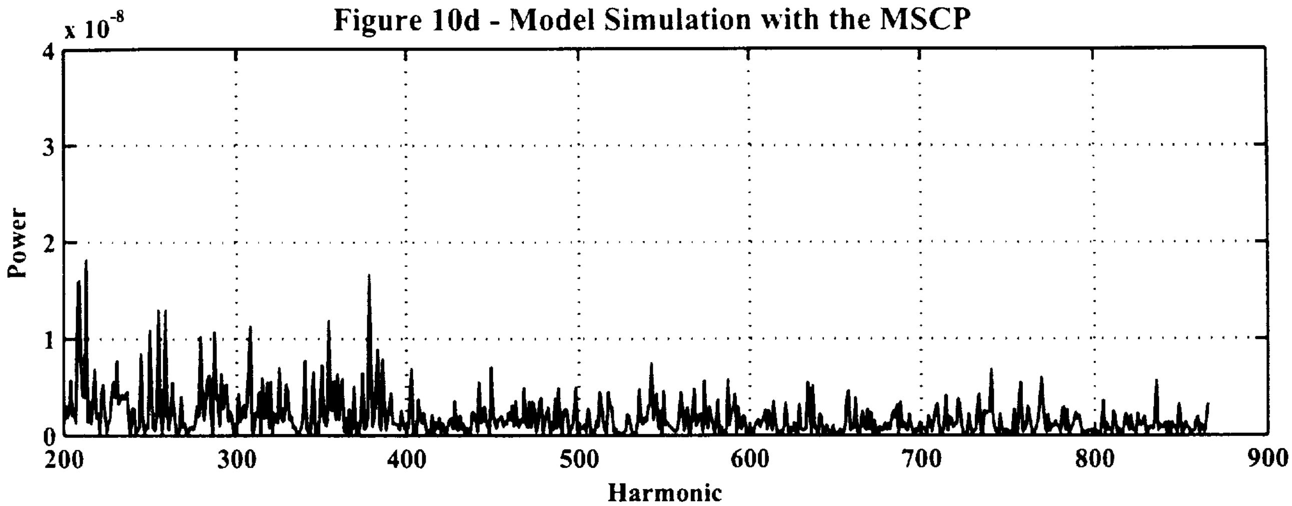
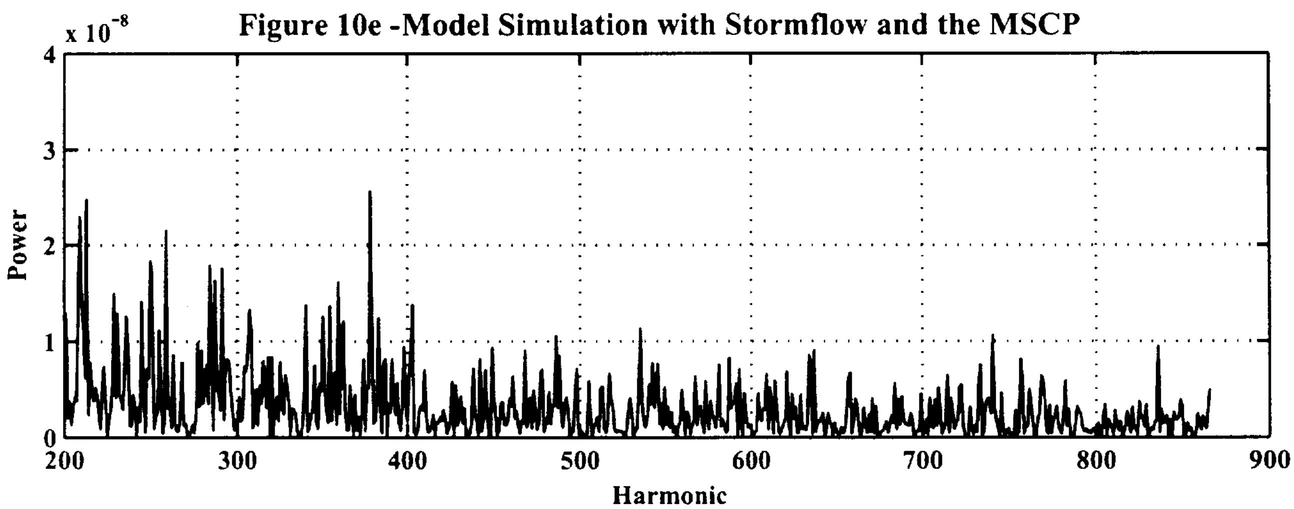


Figure 10c -Model Simulation with Stormflow





x 10^-8
Figure 10d - Model Simulation with the MSCP
Power
0
1
2
3
4
200
300
400
500
600
700
800
900
Harmonic



x 10^-8
Figure 10e -Model Simulation with Stormflow and the MSCP
Power
0
1
2
3
4
200
300
400
500
600
700
800
900
Harmonic